Chapter 7

The Basin and Range Province

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7.1. Introduction

The Basin and Range province is a broad, highly extended terrane embedded within the North American western Cordillera that extends from Canada, through the western United States, and across much of Mexico. The province roughly occupies the space between the Cascade Ranges and Rocky Mountains in the north and the Sierra Nevada and the Colorado Plateau in the middle, and it engulfs the Sierra Madre Occidental Range in northern Mexico (Fig. 7–1). Seismicity, high heat flow, and recent basaltic volcanism indicate that the Basin and Range province is actively extending. Its descriptive name is derived from a particular (and most recent) mode of extensional block-faulting that left the characteristic pattern of alternating basins and ranges across the province (e.g., Gilbert, 1928). This pattern is especially prominent in the Great Basin, a region of internal drainage that occupies most of the northern Basin and Range province. Estimates of the total crustal extension across the Basin and Range province converge to between 50 and 100% (e.g., Hamilton and Myers, 1966; Zoback et al., 1981; Wernicke, 1992), though it is recognized that this extension is not uniformly distributed across the province, but instead occurs as extreme extension (100–300%) in some areas, and as minor extension (<10%) in others.

The Basin and Range is unlike most continental rifts because of the breadth of extended lithosphere there. At its widest points, the Basin and Range province is more than 900 km across, much wider than continental rifts such as the 100–300 km wide Rio Grande (Chapter 6) and East African (Chapter 5) rifts. Other broadly extended regions of crust are observed, but are usually found beneath sea level, as in the case of the extended crust beneath the North Sea, the Bering Strait, and the South China Sea, or beneath thick basinal sedimentary rocks. In contrast, the northern Basin and Range stands at an average 1.5 km above sea level. Much of the Basin and Range province lies within a rain shadow behind the Sierra Nevada and Cascade Ranges, and the resulting dearth of erosion and sedimentation has preserved excellent exposures of extended crust. Good exposures, combined with the enigmatic nature of actively extending continental crust residing at high elevations, has invited a great deal of study; as of 1992 more than 7200 papers and books have been written under a subject heading of "Basin and Range". An unfortunate sampling bias in many types of geologic and geophysical data towards the Basin and Range province north of the Mexico-United States border causes much of the focus of this chapter to be on the United States Basin and Range, where more is known. Exceptions to this are included where possible, such as the more comprehensive study of Mexican Basin and Range crustal and mantle xenoliths.

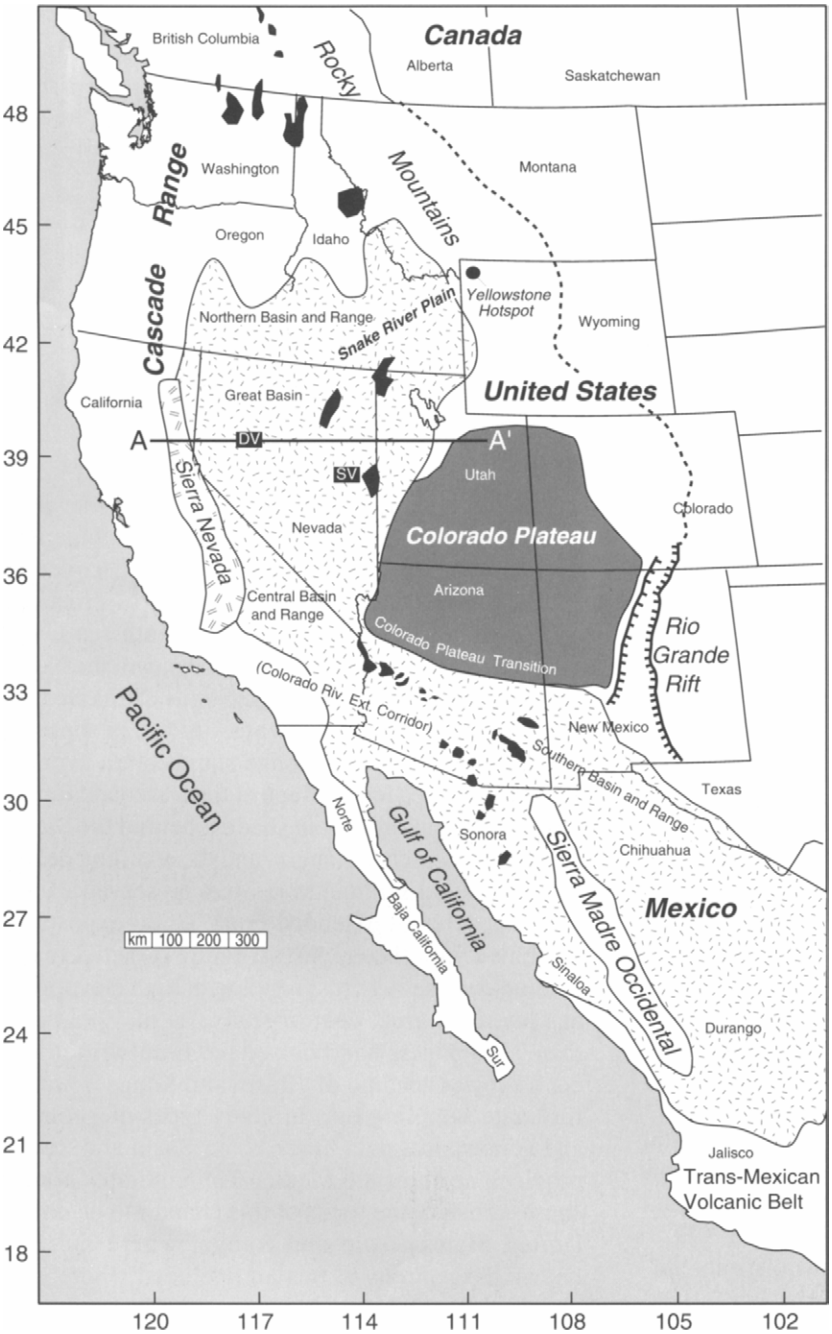


Fig. 7–1. Location of the Basin and Range province (stippled pattern) in relation to other tectonic elements of the North American western Cordillera. The Rio Grande Rift is treated as a separate entity in this study, though is often included as part of the Basin and Range in other regional studies. The black areas are regions of metamorphic core complexes (highly extended terranes; usually on apparent low-angle normal faults), and the patterned area marks the outline of the area traditionally known as the Basin and Range province because of the signature block-faulted topography. Note that a significant part of the Basin and Range area is found in Mexico. A to A' is the location of the cross section shown in Figure 7–15.

7.1.1. Tectonic Cycles in Western North America

Basin and Range extensional structure overprints a long history of tectonism that affected various parts of the western Cordillera (e.g., Conde, 1982). The region has been the locus of many cycles of extensional, compressional, and transform deformation. Late Proterozoic extension created a passive margin along western North America, opening a proto-Pacific basin. This stable continental margin may have persisted until late in the Ordovician Period, when island arcs began to form as a result of convergence and subduction. By Middle Devonian and Mississippian time, the Antler orogeny caused compressional deformation in western North America. By Permian time, back-arc extension caused the formation of linear basins west of the Antler orogenic belt. The Mesozoic Era brought changes in the North American plate drift direction, assembling island arcs together, and moving the trench westward. The Sonoma orogeny was active in the Great Basin, and the great thrust sheets of the Sevier fold and thrust belt that extend from Alaska to Mexico were formed as a result of subduction-generated compressional stresses. By Cretaceous time, an increase in plutonic and volcanic activity created batholiths like those that form the Sierra Nevada. An apparent flattening in the angle of subduction during Late Cretaceous time may have initiated the Laramide orogeny that involved cratonic basement rocks well inboard of the fold and thrust belt, and persisted into the Cenozoic Era, ending abruptly at about 40–50 Ma. "Basin and Range" has become an inclusive term that covers most of the extended crust found in the western Cordillera. A variety of extensional styles were active at various times during the Cenozoic Era, which left their marks across the region, and have come to define the extended province. In general, Cenozoic extensional tectonism is most widely observed in areas of pre-Cenozoic tectonic deformation (e.g., Wernicke, 1992). Middle Cenozoic to Quaternary transform deformation along the western continental margin is partly expressed within the Basin and Range province and links have been suggested between the two deformational patterns (e.g., Zoback et al., 1981; Pezzopane and Weldon, 1993).

7.1.2. Timing and Styles of Basin and Range Extension

The most current interpretation of the Basin and Range province classifies all extensional phases from Eocene time to present as "Basin and Range" extension (e.g., Wernicke, 1992). Two distinct extensional styles are commonly observed across the province: (1) an initial stage of isolated highly extended terranes such as metamorphic core complexes, (regions where mid-crustal rocks are exposed at the surface, exhumed by low-angle normal faults, uplift, and erosion), and (2) a second, later stage of higher-angled block faulting. Highly extended terranes were formed in British Columbia, Canada, and in Washington, Oregon, Idaho, and northern Nevada in the United States during Eocene time. Not all of the Eocene highly extended terranes can be classified as metamorphic core complexes, but they do tend to be localized areas of large-magnitude extension bounded by areas that were not as strongly extended (see Wernicke, 1992, and Axen et al., 1993, for detailed descriptions and locations of these terranes). During Oligocene time, highly extended terranes formed in the Great Basin of eastern Nevada on low-angle detachment faults that may have formed in a back, or intra-arc setting (e.g., Zoback et al., 1981), and clustered along the Proterozoic North American continental edge (e.g., Coney, 1980). Further south, in the southern Basin and Range, the earliest stages of extension began by latest Oligocene time in the southern parts of California and Arizona in the United States, and in Durango, Chihuahua, and Oaxaca, Mexico. By early Miocene time, strong extension had begun on major normal faults across much of Mexico (e.g., Henry and Aranda-Gomez, 1992), and metamorphic core complexes were forming along the Colorado River between California and Arizona (e.g., Howard and John, 1987) and along the southern edge of the Colorado Plateau in southern Arizona (e.g., Rehrig and Reynolds, 1980). A narrow zone of extension between the southern Sierra Nevada and southern Colorado Plateau, sometimes called the central Basin and Range (e.g., Jones et al., 1992; Wernicke, 1992), became active during late Oligocene to early Miocene time, and some of the latest forming metamorphic core complexes are found in this zone (e.g., Wernicke et al., 1988; Axen et al., 1992). The middle Miocene brought the regionally broadest stage of extension to the northern and southern Basin and Range provinces at about 10 and 13 Ma, respectively (e.g., Zoback et al., 1981). This late stage of relatively small magnitude extension on steeply dipping normal faults caused the characteristic block-faulted basin-range structure that gives the province its name. Widespread block-faulting occurred across much of the western Great Basin and southern Arizona, and the opening of the Gulf of California accompanied normal faulting in Mexico at about 12–10 Ma (e.g., Stock and Hodges, 1989). Pliocene and Quaternary eruptions accompany incipient rifting in the Jalisco block that lies at the southern edge of the Sierra Madre Occidental in Mexico (Wallace at al., 1992) (Fig. 7–1), possibly indicating that the Basin and Range province is growing to the south. Large-magnitude earthquakes shake the Basin and Range province occasionally, and are distributed along its entire length, indicating that extension is widespread and ongoing.

7.2. Significance of the Basin and Range Province in reference to worldwide rifting

Two important questions about the Basin and Range province are: why is so much of the province so high, and why is it so wide? Extension generally leads to isostatic subsidence, yet the northern Basin and Range is a high plateau. Moreover, as can be seen from examination of this book, the vast majority of continental rifts are narrow features, and usually form as one or two chains of elongate basins. The Basin and Range province is made up of hundreds of basins of varying depth, age, and orientation that stretch across several hundred km. The origin of the extensional stresses that caused the straining of the Basin and Range is also subject to question, as is the partitioning of extensional strain. For example, why didn't a narrow rift develop into an ocean basin as a result of strong extensional stresses instead of broadly distributed strain? Does extension of orogenic lithosphere manifest itself differently from extension of cratonic lithosphere? A better understanding of the differences between the Basin and Range province and other continental rift zones may help to explain the conditions that lead to more typical rifting, as well as being important to studies of continental extension.

Internal differences in extensional style within the Basin and Range provide further problems as well as possible clues about the extensional process. For example, the northern part of the province stands one km higher in elevation on average than the southern part (Plate 7–1), even though the crustal thickness and composition are largely similar between the two sub-provinces. This difference is a probable indication of a strongly varying, regionally dependent mantle role in Basin and Range extension. The thickness of the crust is generally uniform across the province despite highly variable surface extension. However, the crust is thinner than adjoining unextended provinces like the Colorado Plateau, Rocky Mountains, and the Sierra Nevada, an indication that crustal pure shear does occur, but on a much broader scale than that of individual basins within the extended province. At an average 30 km thick, the 50–100% extended Basin and Range crust must have either started out thick (>45–60 km) (e.g., Hamilton, 1987), or its thickness was inflated during extension by magmatism and/or lower-crustal flow from outside the province. Magmatism is ubiquitous in Basin and Range extension, and the different patterns of Cenozoic volcanism in the Basin and Range province allow for the cataloging and assessment of the tectonic role of magmatic input into extending lithosphere. Identification of apparent low-angle normal faults in the Basin and Range, whose formation seems to have been confined to particular time intervals in different parts of the province, continues to cause controversy as to whether those faults formed at low angles, or evolved into that orientation. The upper-crustal response to extensional stress changed significantly to more numerous, higher-angle faults during the latest stage of extension. The reasons for this change remain unknown, though many ideas abound. In this chapter I attempt to summarize the current state of knowledge about the Basin and Range province, as well as models for its behavior.

7.3. Geologic Observations

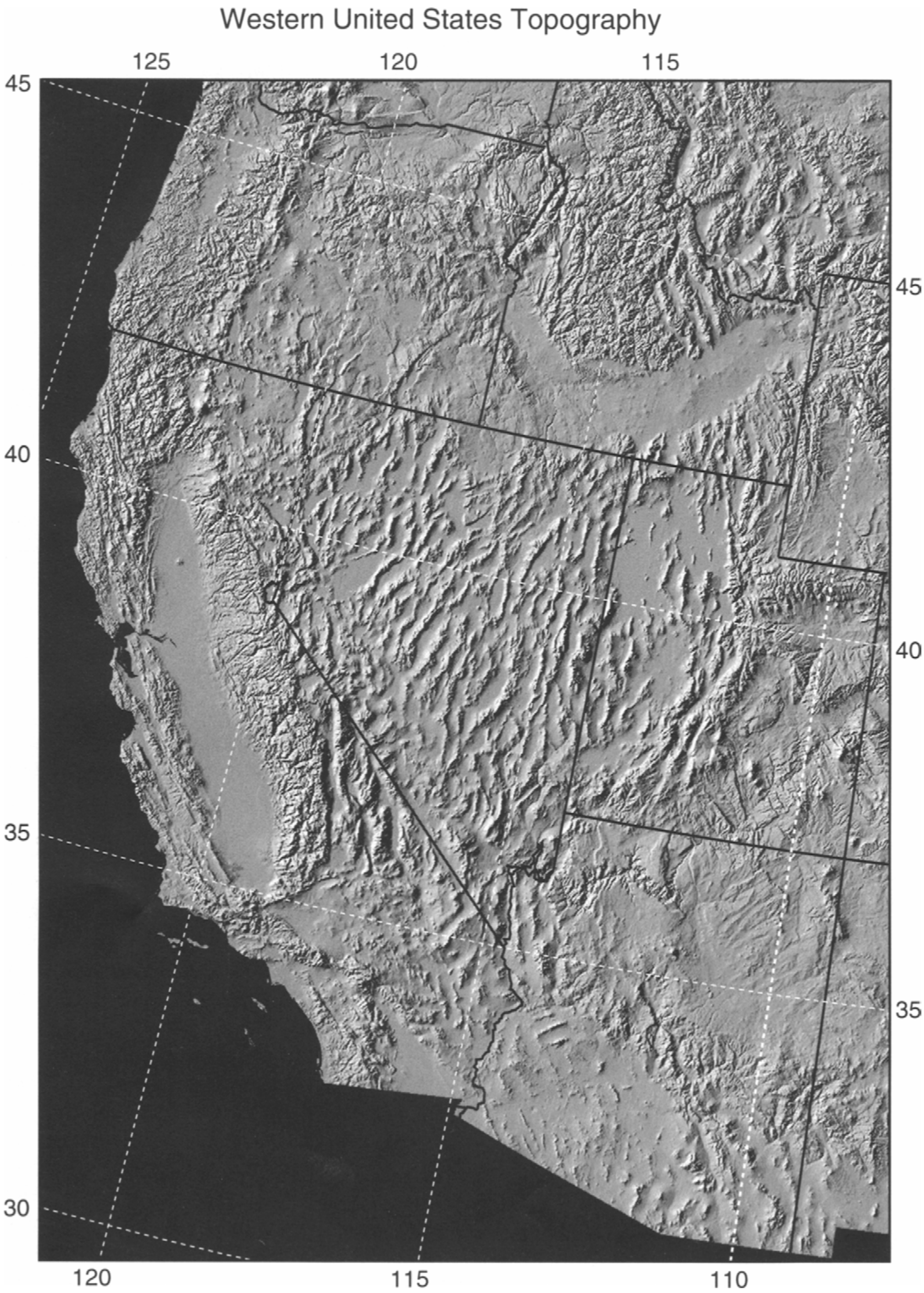
7.3.1. Topography

At an average 1.5 km above sea level, the northern Basin and Range province is unusually elevated for a highly extended province. In general, lithospheric thinning leads to isostatic subsidence, since the buoyant crustal layer is thinned (e.g., Lachenbruch and Morgan, 1990). The southern Basin and Range behaves more typically, and is on average 1 km lower in elevation, with some isolated localities below sea level. The complex tectonism associated with the formation of the continental margin, and later subduction-related orogenies created many inherited topographic features still resolvable in the Basin and Range. The Proterozoic rift that created the western North American continental margin left a thick sequence of sedimentary rocks still present in the eastern Great Basin (e.g., Stewart and Poole, 1974). The Devonian to Mississippian Antler orogeny left an apparent structural division that bisects the northern Basin and Range province into two roughly symmetric sub-provinces (e.g., Eaton et al., 1978). The Sevier fold and thrust belt marks the eastern extent of the Great Basin, but is not observed in the southern Basin and Range. Laramide deformation apparently occurred mostly east of the present-day Basin and Range province, though Laramide structures may be masked by Basin and Range extension (e.g., Dickinson and Snyder, 1979).

Pervasive extension-related topographic variation within the Basin and Range are present at many different scales. The stage of block-fault extension that began in the middle Miocene left behind a series of alternating ranges spaced an average 25–35 km apart, with basins about 10–20 km wide in-between. The typical length-to-width ratio of the ranges is about 4:1 up to 8:1 (Zoback et al., 1981). These basin-range pairs are most pronounced in the Great Basin, but similar structural trends are observed in the southern Basin and Range as well (Fig. 7–2), with the pattern extending across much of Mexico. Block faulted topography overprints the earlier phases of low-angle faulted, domed uplift of mid-crustal rocks (metamorphic core complexes; outlined on Figure 7–1). The eastern Snake River Plain cuts a low topographic swath across the northern Basin and Range (Figs. 7–1,7–2), and marks the track of the Yellowstone hotspot (e.g., Morgan, 1972; Armstrong et al., 1975, Pierce and Morgan, 1992) (Fig. 7–2). Surrounding the low topography of the Snake River Plain is a rugged, actively extending part of the northern Basin and Range province that may be strongly influenced by the Yellowstone hotspot (e.g., Anders and Sleep, 1992; Pierce and Morgan, 1992) (Fig. 7–2). There is a fairly abrupt topographic step between the elevated northern Basin and Range, and lower southern Basin and Range that is located in southern Nevada—approximately between the southern end of the Sierra Nevada and southern Colorado Plateau (Plate 7–1, Fig. 7–2). Like most rifts, much of the Basin and Range province is bounded by high-standing margins, the Sierra Nevada on the west flank, and the Rocky Mountains and Colorado Plateau on the east flank. Seismicity along the margins of the Basin and Range indicates that the rift-flanks may still be rising. The reader is directed to the summary of Mayer (1986) for more detailed discussion of Basin and Range topographic variation.

7.3.2. Sedimentary Record – Two Example Basins

Because of the large number and diverse nature of the basins in the Basin and Range province, the complete sedimentary history of the Basin and Range is beyond the scope of this review. Thus I choose two structurally different, characteristic styles of basin to discuss briefly here; the Spring Valley in eastern Nevada ("SV" on Figure 7–1), which formed during Oligocene time and is associated with the formation of the Snake Range metamorphic core complex, and Dixie Valley in western Nevada ("DV" on Figure 7–1), which formed during later stage middle Miocene high-angle normal faulting. The following summary of Spring Valley is based on studies by Miller et al., (1983), Bartley and Wernicke (1984), Gans et al., (1985), and McCarthy (1986). Spring Valley is located within the Mesozoic Sevier orogenic zone in eastern Nevada, but the Paleozoic section was locally only mildly deformed prior to Tertiary extension. Oligocene normal displacement on the low-angle Snake Range decollement thinned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the Paleozoic section, opened the Spring Valley basin, and elevated Precambrian basement rocks that now core the Snake Range to the east. A well on the east side of the valley penetrated about 1650 m of Tertiary and Quaternary sedimentary rocks overlying 150 m of Paleozoic rocks separated from Precambrian quartzite by the detachment fault at about 1800 m depth. Either later-stage, higher angle normal faulting (Miller et al., 1983; Gans et al., 1985), or rotation of hanging wall blocks on higher angle normal faults that sole into a second detachment beneath the Snake Range decollement (Bartley and Wernicke, 1984), caused a deepening of the western Spring Valley, which contains 3–4 km of Tertiary and Quaternary sediments above a thicker 2 km sequence of Paleozoic rocks.

In contrast, Dixie Valley in western Nevada was formed during the later stage of extension in the Great Basin, when higher-angled block-faulting predominated. The following summary is based on studies by Speed (1976), Thompson and Burke (1973), and Okaya and Thompson (1985). Dixie Valley is an asymmetric, westward-thickening half graben that is controlled on its western side by a ~50° dipping (down to the east) normal fault. Alluvial fan deposits shed from the Stillwater Range fill the upper, western part of the valley, while the rest of the upper section is occupied by a 1-km-thick section of Tertiary lacustrine and playa deposits. Beneath those deposits lies a 500-m-thick section of Tertiary volcaniclastic rocks that rests on Mesozoic basement rocks. Gravity and seismic data show a 3-km total basin depth. Offsets in 12,000-year-old lake-shoreline markers constrain about 1 mm/yr of extension during that time interval, and offsets in an 8-m.y.-old basalt flow constrain a 0.4 mm/yr average extension rate during the past 8 m.y. Large normal-fault earthquakes have occurred as recently as 1954, and continuous microseismicity indicates that extension continues in Dixie Valley. Ongoing deformation has uplifted Tertiary and Quaternary sediments in many places. They preserve a record of the earliest interruption of through-going drainage by block-faulting and formation of basins (Stewart, 1980).

Fig. 7–2. Topography of western North America (Thelin and Pike, 1991) showing the distinctive patterns of block faulting that pervade both the northern and southern Basin and Range province. The eastern Snake River Plain can be seen clearly as a low topographic area in southern Idaho.

7.3.3. Igneous Activity

Synextensional magmatism is nearly always observed as a phase of any Basin-and-Range extensional system. The particular manifestation of magmatism is highly variable across the province: synextensional plutonism from silicic to mafic compositions, dike and sill intrusions of variable compositions, rhyolite and basalt flows, ash flow tuffs, cinder cones, and in some areas, flood basalts are seen as eruptive surface features. The exact relational timing of extensional faulting and associated magmatism is highly variable across the Basin and Range province, and can change with extensional style, timing, and locality. Taylor et al. (1989) found in a transect across the eastern Great Basin that only one of four distinct extensional faulting episodes correlated exactly with nearby volcanic activity, and Best and Christiansen (1991) concluded that extension during peak volcanism in the Great Basin was limited. In contrast, Gans et al. (1989) found cross-cutting relations in the eastern Great Basin indicating that voluminous magmatism slightly predated extension, compared those relations with other parts of the Basin and Range, and concluded that magmatism plays an active role in extension. A close association in space and time between magmatism and metamorphic core complex development has been noted (e.g., Coney, 1980; Glazner and Bartley, 1984; Ward, 1991; Axen et al., 1993). Virtually all examples of low-angle normal faulting within the Basin and Range province, as well as worldwide, have some form of lower-plate intrusive magmatism that can be related to the extensional episode that caused the faulting (Table 7–1). The ubiquitous association of magmatism with low-angle faulting and metamorphic core-complex development has led some authors to suggest a mechanical tie between the processes (e.g., Lister and Baldwin, 1993; Parsons and Thompson, 1993). Further discussion on the mechanics of low-angle normal faulting can be found in Section 7.5.1.

Table 7–1

Core Complex	Associated Lower-Plate Magmatism	Reference
Whipple Mountains- S. Calif.	Mid-Tertiary dike swarms and mafic plutons	(e.g., Davis et al., 1982)
Chemehuevi Mountains- S. Calif.	Mid-Tertiary dike swarms	(e.g., John, 1982)
Homer Mountain-Sacremento Mountains-S. Calif.	Mid-Tertiary dike swarms	(e.g., Spencer, 1985)
Castle Dome Mountains-S. Calif.	Mid-Tertiary dike swarms	(e.g., Logan and Hirsch, 1982)
Black Mountains - S. Calif.	Miocene mafic dikes and plutons	(e.g., Asmerom et al., 1990)
Eldorado Mountains- S. Calif.	Tertiary Mafic-silicic dikes and plutons	(e.g., Anderson, 1971)
Mopah Range - S. Calif.	Cross-cutting Tertiary mafic dikes/low-angle faults	(e.g., Hazlett, 1990)
Buckskin-Rawhide Mountains - Arizona	Tertiary mafic-silicic plutons	(e.g., Bryant and Wooden, 1989)
Santa Catalina-Rincon-Tortolita - Arizona	Mid-Tertiary granite plutons	(e.g., Kieth et al., 1980)
South Mountains - Central Arizona	Mafic and Silicic Mid-Tertiary dikes and plutons	(e.g., Reynolds and Rehrig, 1980)
Pinaleno-Santa Teresa Mountains - Arizona	Mid-Tertiary silicic dike swarms	(e.g., Rehrig and Reynolds, 1980)
Harcuvar Mountains - Arizona	Tertiary Pluton	(e.g., Rehrig and Reynolds, 1980)
Harquahala Mountains - Arizona	Mid-Tertiary silicic dikes and pluton	(e.g., Rehrig and Reynolds, 1980)
Mojave Mountians - W. Arizona	Upper plate (?) Mid-Tertiary mafic-silicic dike swarms	(e.g., Nakata, 1982)
Newberry-Dead Mountains- S. Nevada	Tertiary Mafic-Silicic dikes	(e.g., Simpson in prep.)
Ruby Range - NE Nevada	Mid-Tertiary deep structural granitic intrusions	(e.g., Wickham et al., 1993)
Snake Range - Nevada	Oligocene-Miocene Mafic dikes	(e.g., Lee et al., 1987)
Kettle - Okanogan Domes -E. Washington	Eocene Mafic dike swarms	(e.g., Holder et al., 1990)
Mykonos, Aegean Sea - Greece	Miocene dikes - plutons	(Lee and Lister, 1992)
Cyclades Islands, Aegean Sea - Greece	Coeval magmatism	(Lister et al., 1984)
D'Entrecasteaux Islands - Papua New Guinea	Granodiorite plutons coeval with low-angle detachment	(Hill et al., 1992)

Table 7-1. Compilation of metamorphic core complexes and observed mode and best timing constraints of magmatism in relation to low-angle faulting. Virtually all core complexes have some form of lower-plate intrusive magmatism associated with them.

Because of the huge volume of magmatic activity across the Basin and Range province, I focus here on the broad space-time patterns of magmatic activity that have swept across the province during Tertiary time. The following discussion is based primarily on reviews and studies by Coney and Reynolds (1977), Burke and McKee (1979), Lipman (1980), Zoback et al. (1981), Gans et al. (1989), Armstrong and Ward (1991), Henry and Aranda-Gomez (1992), Jones et al. (1992), and Axen et al. (1993). At the beginning of Tertiary time, magmatism occurred north and south of what is called the Laramide gap

(Fig. 7–3a), an interruption in arc-related magmatism that extended from the middle of Nevada and Utah to the United States-Mexico border, and initiated at about 80 Ma. This gap is thought to relate to low-angle subduction of the Farallon slab, which may have cooled the asthenospheric wedge where melt was generated (e.g., Dumitru et al., 1991), and minor volcanism was shifted eastward into the Rocky Mountains. The beginning of Tertiary time was also associated with the culmination in a decline of magmatic activity that began at 80 Ma. From 65 to 60 Ma, a rapid increase in activity swept through the region north of the Laramide gap, and the beginnings of Eocene extension in southwest Canada and the northwest United States brought increased magmatism between 60 and 55 Ma.

During Eocene time, a gradual southern sweep of extension-related magmatism worked its way from Idaho and Montana south into northern Nevada and Utah. These melts were typically of andesitic and rhyolitic compositions, and probably resulted from heavy contamination of mantle-derived basalts. South of the Laramide magmatic gap in Mexico, the volcanism that occurred during Eocene time was probably subduction-related. By late Eocene and early Oligocene time, a strong shift in extensional magmatic activity to the southern Great Basin occurred, centered in Nevada and Utah, and is referred to as the Great Basin ignimbrite flare-up (Fig. 7–3a). The Laramide gap persisted through Oligocene time, separating an episode of magmatism that swept westward across southeast California and southern Arizona. Independent centers of activity in southern Colorado and eastern Arizona were joined by magmatism associated with the opening of the Rio Grande rift (Fig. 7–3a).

The late Oligocene and early Miocene brought a northward sweep of core complex extension and magmatism from Mexico into the Colorado River extensional corridor, and a closing of the Laramide magmatic gap. The joining of the Nevada and Arizona magmatic centers was associated with a change from primarily intermediate composition melts into a primarily bimodal system predominated by basaltic eruptions. By middle Miocene time, basaltic volcanism dominated most of the active Basin and Range province from Mexico to northern Nevada. A westward shift in magmatic activity, correlated with incipient extension in the western Great Basin, occurred at this time, and a northeastward shift from southern Arizona across the southwest Colorado Plateau margin was also associated with the onset of extension there (Fig. 7–3b). During this period, magmatism appears to have moved into stable areas just prior to the onset of extensional faulting. The Yellowstone hotspot broke through in northern Nevada during middle Miocene time (~17–16 Ma), and created a series of northeast-younging caldera systems as the North American plate tracked over the plume. Extensive flood basalts followed the earlier stages of caldera formation; these basalts continue to erupt across much of the plume track (eastern Snake River plain, Idaho). Formation of the northern Nevada rift, a 500-km-long magmatic feature that extends south from eastern Oregon to southern Nevada, is thought to have resulted from dike injection from the Yellowstone plume (Zoback and Thompson, 1978); and it is possible that the Columbia Plateau flood basalt province is related to the Yellowstone plume as well. The extent to which the Yellowstone plume has affected the extensional magmatic patterns in the Basin and Range province (if at all) remains controversial, and will be further addressed in later sections. Magmatic activity from late Miocene time to the present has migrated away from the regions of strongest middle Miocene activity in central Nevada and southern Arizona, and has concentrated at the outer margins of the Basin and Range province (Fig. 7–3c).

7.3.4 Xenolith Studies

Xenoliths are found in the northern Basin and Range province in the eastern Snake River Plain in Idaho, and in the Great Basin at one locality in central Nevada. In the southern Basin and Range xenoliths are more abundant; localities include Mexico, southern California, and southern Arizona. On the eastern Snake River Plain, Archean lower-crustal xenoliths were brought to the surface by Quaternary basalts, possibly indicating that Archean crust underlies that region (Leeman et al., 1985). Menzies, in a 1989 study that includes xenoliths from central Nevada divides the western United States into mantle

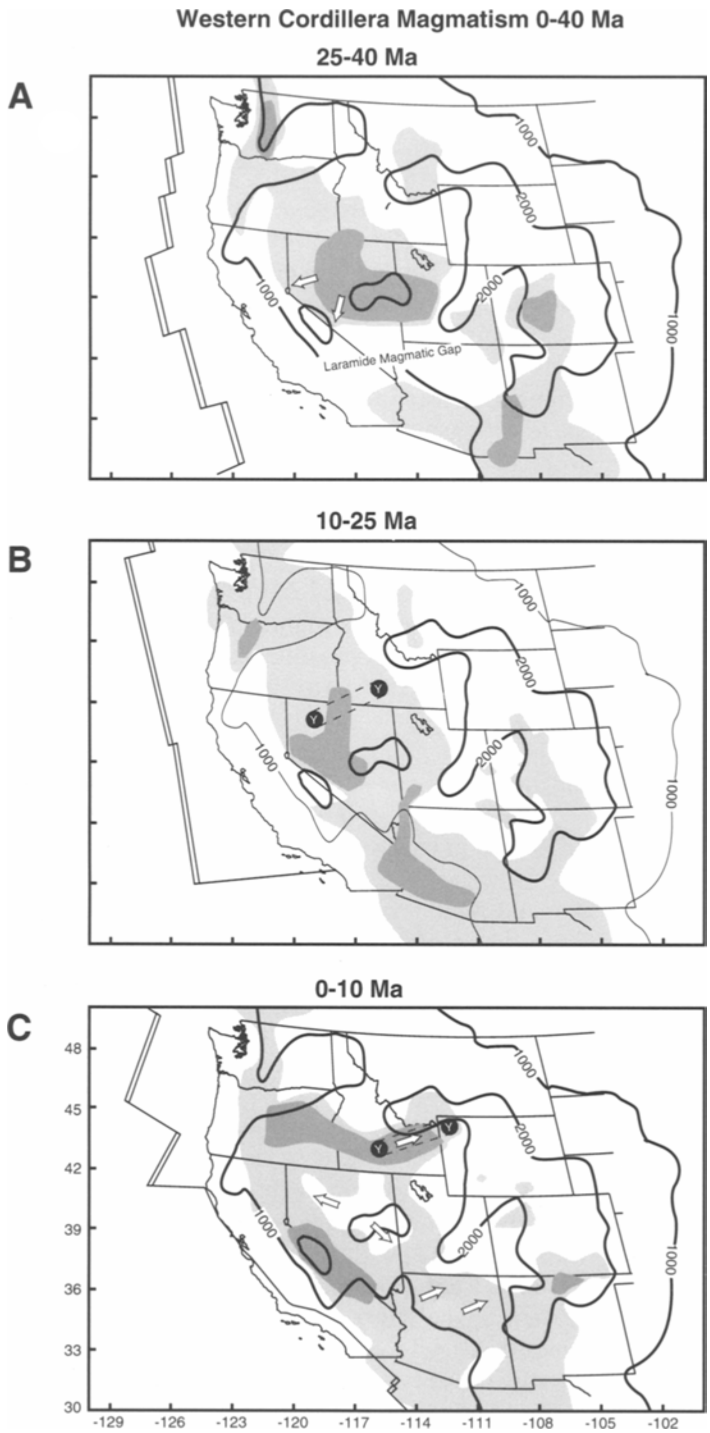

Fig. 7–3. Space-time patterns of magmatic activity in the western United States since Eocene time (after Armstrong and Ward, 1991). The darkest areas correspond to peak activity during the time interval. Large arrows on the continent show the approximate directions of magmatic migration during the time interval, and the smaller arrows show plate motion in the hotspot frame. The approximate locations and motion directions of the subducting Farallon plate are also shown. The inferred location of the Yellowstone plume is shown in windows (b) and (c). Details on the magmatic patterns are discussed in the text. Contour lines are smoothed topography in meters.

domains based on seismic tomography, heat flow, and xenolith thermobarometry. He places the Basin and Range province within an oceanic mantle domain, because of similarities of xenolith isotope-signature to mid-ocean and ocean-island basalts. Upper mantle and lower crustal xenoliths are found in southern California and southern Arizona in Tertiary volcanic rocks. The upper mantle xenoliths are lherzolites and harzburgites, and commonly show evidence for brittle deformation related to multiple stages of intrusive events in the upper mantle and also ductile deformation possibly related to tectonic extension (Wilshire, 1990; Wilshire et al., 1991). Composite xenoliths containing two or more rock types indicate that the crust-mantle boundary is very likely interlayered and has evolved through multiple stages of magmatic intrusion (Wilshire et al., 1991). Lower-crustal xenoliths are also found in southern California and Arizona and consist of mafic and ultramafic gabbros and microgabbros of intrusive origin, indicating that magmatic intrusion has also shaped the lower crust in the region as well as the upper mantle (Wilshire, 1990; McGuire, 1992).

Studies of xenoliths found in Mexico consistently report high-temperature granulite facies rocks from the lower crust and upper mantle beneath northern and central Mexico (Nimz et al., 1986; Ruiz et al., 1988; Hayob, et al., 1989; Roberts and Ruiz, 1989; Rudnick and Cameron, 1991). Upper mantle xenoliths are primarily spinel-lherzolites, and lower-crustal xenoliths are primarily granulites and banded gneisses. All the xenoliths are thought to originate from very near the crust-mantle boundary. Very high metamorphic temperatures up to 950–1100 °C (Hayob et al., 1989) are reported, and highly variable ages suggest that this metamorphism has occurred in many stages from 1.1 Ga to as young as 1 Ma in northern Mexico (Rudnick and Cameron, 1991), and includes a post-Oligocene stage in central Mexico (Hayob et al., 1989). The high-grade metamorphism in the xenoliths from Mexico is suggested to be the result of a regional basaltic underplating event that took place sometime after 30 Ma (Hayob et al., 1989). However, Torgersen (1993) finds that low $^3He/^4He$ ratios, while indicating the presence of some mantle-derived magmas, are in general too low for massive underplating to have occurred.

7.3.5 Isotope Geochemistry, Major and Trace Element Studies

The great abundance of igneous rock generated during Basin and Range extension has led to extensive study of the chemistry and composition of these igneous rocks and inclusions. The isotope and geochemical characteristics of mantle-derived melts can reveal clues about the mantle and crustal columns through which they intrude. Variable lithospheric-mantle ages and compositions can be deduced from isotope ratios in neodymium, strontium, and lead. The origins of basalts from different regions within the Basin and Range province vary from asthenospheric sources with minor lithospheric contamination, to those that include significant components of melted lower crust and/or lithospheric mantle. Glazner and Farmer (1991) point out that subtle crustal contamination by mafic crust can be mistaken for subcontinental mantle-source variability in some instances. However, there is good agreement in the broad isotopic patterns observed across the Basin and Range province and margins, which can be gleaned from the variety of studies of extension-related igneous rocks from the Basin and Range province presented below.

Many researchers have noted extreme variability in the western United States subcontinental mantle based on strontium and neodymium isotope variation. Menzies et al. (1983) suggested that melts from the Basin and Range province tapped a source similar to mid-oceanic ridge basalts, while melts from the Snake River Plain, and Sierra Nevada retained the primary isotopic characteristics of the underlying mantle lithosphere. Lum et al. (1989) tested possible contamination modes against the observations that Snake River Plain magmas have apparent lithospheric sources while central Nevada magmas appear to have asthenospheric sources against possible contamination modes; they concluded that contamination cannot account for the variations and that upwelling asthenosphere beneath the highly extended Great Basin causes the observed differences.

Leeman and Harry (1993) proposed two stages of magmatism in the Great Basin; the first, from 40–5 Ma involved melting of Precambrian-aged mafic veins and pods within the mantle lithosphere, and the second (since 5 Ma) involved melting of upwelling asthenosphere. Fitton et al. (1988) also suggested an asthenospheric source for late Cenozoic Basin and Range basalts, and further note a bilateral symmetry in basalt chemistry from the Great Basin that mirrors the topographic and gravity signatures pointed out by Eaton et al. (1978). They also noted increases in lithospheric-mantle involvement in basalts erupted around the outer margins of the Basin and Range. Regional lower-crustal cooling of about 300 °C from early Oligocene to early Miocene time was concluded to have caused reduced crustal contribution to rhyolite eruptions as inferred from neodymium isotope data (Perry et al., 1993).

Liviccari and Perry (1993) contoured zones of neodymium-depleted mantle model ages in the western United States, and observed that much of the Precambrian mantle lithosphere beneath the northern Basin and Range province has been preserved, although the dominance of asthenospheric sourced Late Cenozoic basalts in the southern Basin and Range caused them to suggest lithospheric removal there. Ormerod et al. (1988) located a roughly north-south oriented lithospheric boundary associated with strong variation in strontium and neodymium ratios in the western Great Basin, near the Sierra Nevada. They also found a second, northward-younging asthenospheric-source component that can be correlated with the trailing edge of the subducted Juan de Fuca plate. Farmer et al. (1989) analyzed isotope ratios in post-10-Ma basalts, and located a boundary in southern Nevada that separates asthenospheric-source basalts in central Nevada from lithospheric-source basalts in southern Nevada. They suggest that the Laramide magmatic gap may have preserved the mantle lithosphere in southern Nevada, as did Liviccari and Perry (1993). Temporal isotope and bulk chemistry variations were used to constrain the depth of magma generation in the same part of southern Nevada, and 50% less thinning of the mantle lithosphere than would be predicted from surface extension was inferred there (Daley and De Paolo, 1992). Further south, the southern Cordillera basaltic andesite province erupted across northern Mexico and southern Arizona from about 32 to 17 Ma and has a typical arc-like trace element signature; stratigraphic analysis indicates these rocks were emplaced in an extensional environment, perhaps constraining initial extension in Mexico to a back- or intra-arc setting (Cameron et al., 1989).

At about 16–17 Ma the Yellowstone plume emerged, and accompanying basaltic volcanism began to dominate in the northern Basin and Range. Interpretations of Neogene Cordilleran basalt compositions vary as to whether or not they are consistent with a mantle plume source; some workers have suggested that they are (e.g., Fitton et al., 1991; Menzies et al., 1991), whereas others have found that isotopic signatures in places are more closely related to the underlying lithosphere (e.g., Lum et al., 1989; Lipman and Glazner, 1991; Bradshaw et al., 1993). A mantle plume acts more as a source of voluminous hot material than as a point source of heat (e.g., Sleep, 1990), and it is likely that conducted heat applied broadly to the lithosphere by a plume head would cause some melting of the existing mantle lithosphere as well as supply more primitive melts directly from the asthenosphere.

7.4. Geophysical Observations

7.4.1. Seismicity

The Basin and Range province is a seismically active region (Fig. 7–4). Nearly as many historical earthquakes greater than magnitude 7 have occurred in the Basin and Range as have along the San Andreas Fault system (e.g., Ryall et al., 1966; Thompson and Burke, 1974). The 1983 M_s 7.3 Borah Peak earthquake that struck just north of the eastern Snake River Plain in Idaho serves as a reminder that the Basin and Range continues to pose a considerable seismic hazard. The following summary of Basin and Range seismicity is based on studies by Thompson and Burke (1974), Smith (1978), Smith and Lindh (1978), Stickney and Bartholomew (1987), Dewey et al. (1989), Smith et al. (1989), and Pezzopane and Weldon (1993). Seismicity in the Basin and Range tends to occur as episodic swarms or clusters of events that strike previously quiescent

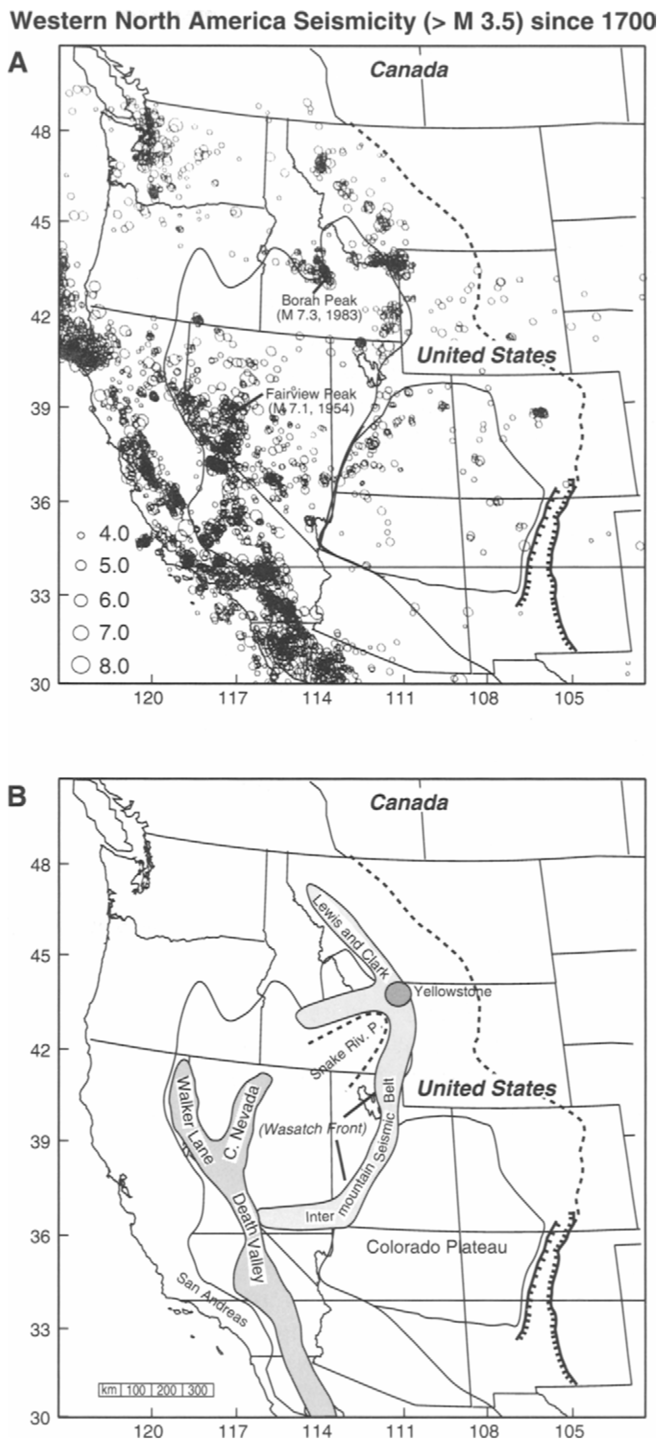

Fig. 7–4. (a) Western North American seismicity (events > M 3.5) since 1700 (after Dewey et al., 1989). (b) The patterns of extensional and oblique seismic strain discussed in the text are outlined in gray below. The dashed line is meant to illustrate the apparent parabolic distribution of seismicity around the eastern Snake River Plain. In general the Basin and Range is most active in the north, especially in the outer margins. Geologic provinces are outlined as indicated on Figure 7–1.

areas, which after some period of activity, return to quiescence. Much like the most recent patterns of magmatism, seismicity in the Basin and Range is most active along its margins (Fig. 7–4). As would be expected in an extensional province, fault-plane solutions show that normal-fault slip dominates, though many oblique and pure strike-slip earthquakes are observed as well. The maximum observed focal depth of earthquakes in the Basin and Range province is about 15 km, although the majority of quakes are shallower. The depth to the brittle-ductile transition is inferred to be at 15 km based on this observation (e.g., Sibson, 1982).

Apart from the widespread, scattered seismicity that occurs across the entire Basin and Range province, there are two broad zones of increased activity along the eastern and western margins of the northern part of the province (Fig. 7–4). Along the eastern margin, the Intermountain seismic belt extends through Utah, eastern Idaho, and western Montana. In southern Utah and Arizona the Intermountain seismic belt is located along the edges and outer margins of the Colorado Plateau. Generally low-level seismicity tends to cluster along the boundaries of the stable Colorado Plateau crustal block, perhaps an indication that its margins are collapsing in extension. To the north, the rising Wasatch Front (Fig. 7–4) marks the boundary between the Basin and Range province and the more stable Rocky Mountains, and has created a zone of active north-south striking subparallel faults. North of the Wasatch Front, the Yellowstone caldera is a center of shallow seismic activity related to the ascent and emplacement of magma in the crust. An apparent parabolic distribution of seismicity distributed around the eastern Snake River Plain in Idaho (e.g., Anders et al., 1989) includes the 1983 Borah Peak earthquake and aftershocks. The interior of the eastern Snake River Plain is virtually aseismic (e.g., Jackson et al., 1993), and it is thought that magmatic strain accommodation of some form may prevent large earthquakes from occurring there (e.g., Parsons and Thompson, 1991; Anders and Sleep, 1992). North of the eastern Snake River Plain, a diffuse zone of intermittent seismicity trends northwest across western Montana, and is associated with a shear zone known as the Lewis and Clark line.

A second zone of seismicity is found along the western boundary of the Basin and Range province. This zone is interrelated to the transform domain along coastal California and Basin and Range extension; oblique slip, transtensional deformation, and extensional transform faulting are commonly mixed with normal faulting along this zone. In the southernmost Basin and Range, seismicity is concentrated along the spreading system in the Gulf of California. Further to the north this seismic belt branches away from the southern end of the San Andreas Fault zone, and crosses southeastern California into Death Valley (Fig. 7–4). The central Nevada seismic belt, consisting primarily of extensional earthquakes, branches northeast away from this zone, and includes recent large earthquakes such as the 1954 M 7.1 Fairview Peak earthquake. The Walker Lane, a complex zone of primarily right-lateral slip, continues in a northwesterly trend along the Sierra Nevada-Basin and Range boundary. Oblique rifting continues along this trend into eastern and central Oregon.

7.4.2 The State of Stress

The state of stress in the Basin and Range province has been determined from earthquake focal mechanisms, hydraulic fracture experiments, recent fault slip indicators, young volcanic alignments, and borehole deformation studies. Some indicators, such as dated volcanic rocks, allow determinations at particular times in the past and focal mechanisms provide information at hypocentral depths. Not surprisingly, the dominant stress state is extensional, with the greatest principal stress oriented near-vertical, the least principal stress oriented broadly east-west, and the intermediate stress oriented broadly north-south (Fig. 7–5). Below, variations and features of the Basin and Range stress field are summarized after studies by Thompson and Zoback (1978), Zoback and Zoback (1980), Zoback (1989), Zoback and Zoback (1989), Mount and Suppe (1992), and Suter et al. (1992). Stress-directions discussed below will be the horizontal component of the least horizontal stress, which is essentially parallel to the extensional strain direction.

Fig. 7–5. Directions of the greatest horizontal stress in the Basin and Range province and Colorado Plateau (after Zoback and Zoback, 1989). The approximate direction of extensional strain is perpendicular to the greatest horizontal stress. The stress directions are variable across the province, changing from west to east. Further discussion can be found in the text.

Stress measurements are sparse in the southern Basin and Range. In central Mexico and along the spreading ridge in the Gulf of California, the least principal stress direction is generally aligned in a northwesterly direction. Along the southern and western edges of the Colorado Plateau, the least-stress directions are aligned radially away from the center of the plateau, a further suggestion that the plateau margin is collapsing gravitationally. The western flank of the Colorado Plateau in Utah shows a northwesterly orientation in the least principal stress (Fig. 7–5), which is parallel to the stresses in the eastern Great Basin, a possible indication of Basin and Range extension encroaching into the plateau. To the north, the margins of the Great Basin show approximately east-west least principal stress directions, whereas the interior of the Great Basin, in the Nevada seismic belt, the least principal stress direction is oriented more northwesterly. North of the Wasatch Front, the least principal stress direction changes again and is oriented more towards the southwest in the region around the eastern Snake River Plain, perhaps as a result of uplift by the Yellowstone plume (e.g., Pierce and Morgan, 1992). In general the variations in stress orientations across the Basin and Range are not well explained, but probably result from variations in crustal rheology and forces applied from outside the domain.

Table 7–2

Strain Rate 1/s	Time Interval	Basis	Location	Reference
3.80E-16 (1cm/800km/yr.)	present	Plate motions and space Geodesy	Northern Basin and Range: Colorado Plateau to Sierra Nevada	1,2,3
3.80E-16 (1cm/800km/yr)	historical earthquakes	Summation of seismic moments	Northern Basin and Range	4,5
6.30E-16 (10.7km/52km/10Ma)	10 Ma +/- 4 my	Lateral and vertical offsets of northern Nevada rift	North-central Nevada	6,7
7.90E-16 (10m/30km/12Ka)	Holocene	Offsets of 12 Ka lake shoreline; 30 km between ridge crests	Dixie Valley, northern Nevada	8
4.10E-16 (3km/30km/8Ma)	8 Ma	Offset of 8 Ma basalt; 30 km between ridge crests	Dixie Valley, northern Nevada	9
2.90E-16 (4m/27km/15Ka)	15 Ka	Two offsets of 15 Ka alluvial surface by Lost River fault; 27 km between ridge crests	Site of 1983 Borah Peak earth-quake N of E. Snake River Plain (Thousand Springs segment)	10
8.20E-16 (~14km/80km/7Ma)	7 Ma (age of calderas longitude Lost Riv. Fault)	Dip of faults taken to be 50 degs. and strata 15 degs. Three tilt blocks spanning 80 km.	Three tilt-block ranges N. of E. Snake Riv. Plain (Lost River, Lemhi, and Beaverhead Ranges, ID)	5,11
11.0E-16 to 17.0E-16 (21-34mm/600km/yr)	16 Ma	Migration rate of rhyolitic volcanism less plate motion rate	Track of hotspot from emergence point to Yellowstone, NV and ID	12

Table 7-2. Northern Basin and Range extensional strain rates from Miocene time to present as derived from a variety of methods. Opening has slowed by about 50% since middle and late Miocene time. References: 1. De Mets et al. (1990), 2. Minster and Jordon (1987), 3. Beroza et al. (1985), 4. Eddington et al. (1987), 5. R. B. Smith et al. (1989), 6. Zoback (1978), 7. Zoback (1979), 8. Thompson and Burke (1973), 9. Okaya and Thompson (1985), 10. Scott et al. (1985), 11. Thompson (1960), 12. Rodgers et al. (1990).

7.4.3 Strain Rate from Geodetic and Other Observations

The present and historical extensional strain rate in the Basin and Range province has been determined from a variety of methods including the tracking of geological markers, space geodesy, and summation of earthquake moments (Table 7–2). The most recent (middle Miocene to present) strain rates are best constrained in the northern Basin and Range and converge to an average 6×10^{-16} s^{-1} strain rate over the entire interval. The present-day rate of about 3.8×10^{-16} s^{-1} is slower than the historical average (e.g., Beroza et al., 1985; Minster and Jordon, 1987; Smith et al., 1989; DeMets et al., 1990). Extensional strain in the Basin and Range has to be considered when reconciling the difference between complete relative Pacific-North American plate motion and San Andreas Fault motion. Argus and Gordon (1991) compared very long baseline interferometry (VLBI) data for motions of sub-elements within the western Cordillera with the NUVEL-1 model (DeMets et al., 1990) for Pacific-North American motion. They found that the San Andreas fault motion differs from ideal Pacific-North American relative plate motion and that Basin and Range extension, westward drift of the Sierra Nevada block, and to a lesser extent, compression across the San Andreas fault absorbs the difference. Clark et al. (1987) determined from VLBI data that Basin and Range extension accounts for 9–10 mm/yr of Pacific-North American plate divergence.

7.4.4. Crustal and Upper–Mantle Structure from Seismic Observations

7.4.4.1. Seismic Reflection Profiling in the Basin and Range Province

Deep, whole-crustal seismic reflection data have been collected in the Basin and Range (Fig. 7–6) province along an east-west transect across Nevada at about the latitude 40°N (e.g., Klemperer et al., 1986; Allmendinger et al., 1987; Jarchow et al., 1993), in the Mojave Desert of southern California (e.g., Cheadle et al., 1986), in Death Valley, California (de Voogd et al., 1986; Serpa et al., 1988; Brocher et al., 1993), across the Whipple Mountain metamorphic core complex in southern California (e.g., Flueh and Okaya, 1989), in Dixie Valley, Nevada (Okaya, 1986), and across the southern Basin and Range and Colorado Plateau transition in southern Arizona (e.g., Hauser et al., 1987; Goodwin and McCarthy, 1990; Howie et al., 1991). Vertical-incidence deep-crustal seismic data from the Basin and Range province tend to show such general features as upper-crustal Cenozoic extensional structures that overprint Precambrian, Paleozoic, and Mesozoic structures, all overlying a highly reflective laminated lower crust. Often, upper-crustal reflectivity is absent, and the crust is transparent until the lower-crustal zone of reflectivity is reached (Fig. 7–7). The Moho is typically a bright, high-amplitude series of reflections and/or an abrupt termination of reflectivity and tends to be relatively flat despite the presence of strongly varying topography and surface extension above it.

In the northern Basin and Range, the 40° N Consortium for Continental Reflection Profiling (COCORP) transect showed that the reflective textures changed from diffuse dipping reflections within the province margins in California and Utah into a strong, sub-horizontal pattern of reflectivity in the extended crust of Nevada (e.g., Allmendinger et al., 1987). Similar changes in reflectivity patterns from unextended into extended crust are noted elsewhere, and it is thus thought that the laminated lower-crustal reflectors are generated during the extension process (e.g., McCarthy and Thompson, 1988). Because of the remarkable continuity in reflection character in the Moho across the northern Basin and Range, and the lack of a correlation of Moho structure with upper-crustal tectonic features along the transect, the Moho boundary is thought to be young and to have evolved during Cenozoic extension (Klemperer et al., 1986). Near-vertical incidence data were collected near the COCORP 40° N transect in 1986 as part of the Nevada Program for Array Seismic Studies of the Continental Lithosphere (PASSCAL) experiment, and an extraordinarily bright, high amplitude Moho reflection was observed beneath Buena Vista Valley in northwestern Nevada. Carbonell and Smithson (1991) suggested that the high-amplitude Moho transition might result from interlayered melt zones, and Jarchow et al. (1993) constrain the event to be a single molten sill no greater than 200 m thick. Extended correlation of industry reflection profiles from Dixie Valley in the northern Basin and Range also shows a reflective lower crust and a layered, transitional Moho boundary (Okaya, 1986). Deep reflection profiles from the central Basin and Range in Death Valley show a bright, mid-crustal reflection that is similar to the mid-crustal melt body beneath the Rio Grande rift (see Chapter 6); this reflector is also interpreted as a melt body (de Voogd et al., 1986). However, bright reflections from the lower crust in the nearby Amargosa Desert in southern Nevada are interpreted as shear zones rather than as melt bodies (Brocher et al., 1993).

Deep-crustal reflection data were collected in the California Mojave Desert in the southern Basin and Range province in 1982 by COCORP. These data have a similar texture to the profiles collected at the Basin and Range margins along the 40° N COCORP transect in that they seem to show many preserved pre-Tertiary structures, though the possibility exists that more recent, deep low-angle normal faults were imaged (Cheadle et al., 1986). Further east, deep-crustal data were collected over the Whipple Mountain metamorphic core complex near the Colorado River in southeastern California. These data show ample mid-crustal reflectivity that is interpreted as mylonitized lower-plate rocks beneath the Whipple Mountain detachment fault extending to depth (Flueh and Okaya, 1989). In 1986 COCORP collected vertical-incidence data along a northeast-directed transect that extended from the Colorado River into

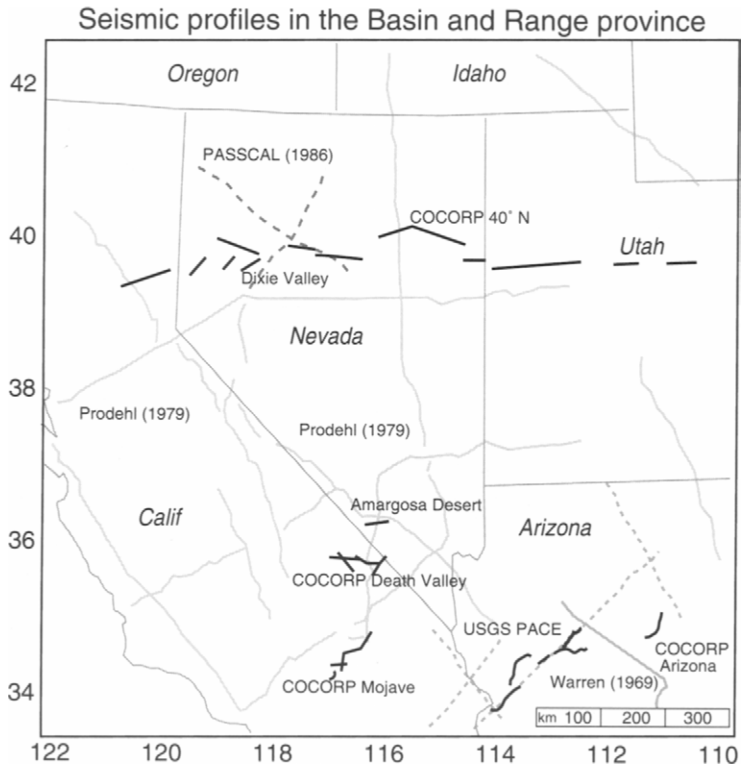

Fig. 7–6. Locations of vertical-incidence and long-offset seismic profiles in the Basin and Range province. Solid black lines are vertical-incidence reflection profiles, while gray lines are long-offset refraction profiles.

the Colorado Plateau of Arizona (Hauser et al., 1987). This transect crossed a variety of extensional provinces from the block-faulted southern Basin and Range, through the Buckskin-Rawhide metamorphic core complex, into the mildly extended Colorado Plateau transition zone (Fig. 7–8). The sections from the southern Basin and Range are similar in appearance to those from the northern Basin and Range in that bright mid-crustal reflectivity is observed above a flat-lying, high amplitude Moho reflection (Hauser et al., 1987). Beneath the Buckskin-Rawhide metamorphic core complex, the mid-crustal reflectivity is fine-scaled and appears disrupted by the extension process (McCarthy and Parsons, 1994). The Colorado Plateau transition zone is actively extending, although the extension is of small magnitude compared with most of the southern Basin and Range. Vertical-incidence data collected in this region by COCORP (Hauser et al., 1987) and Stanford University (Goodwin and McCarthy, 1990; Howie et al., 1991) show preserved intrusive structures of possible Precambrian age in the upper crust, but the middle and lower crust show evidence of recent extensional reworking (Goodwin and McCarthy, 1990; Howie et al., 1991), including a mid-crustal magma body (Parsons et al., 1992a). The reader is directed to Smithson and Johnson (1989) for a more detailed discussion of vertical-incidence studies in the western Cordillera; this discussion also describes numerous shallow to mid-crustal reflection profiles collected in the Basin and Range province that are not discussed here. Also, Mooney and Meissner (1992) compare vertical incidence studies from a variety of extensional provinces worldwide.

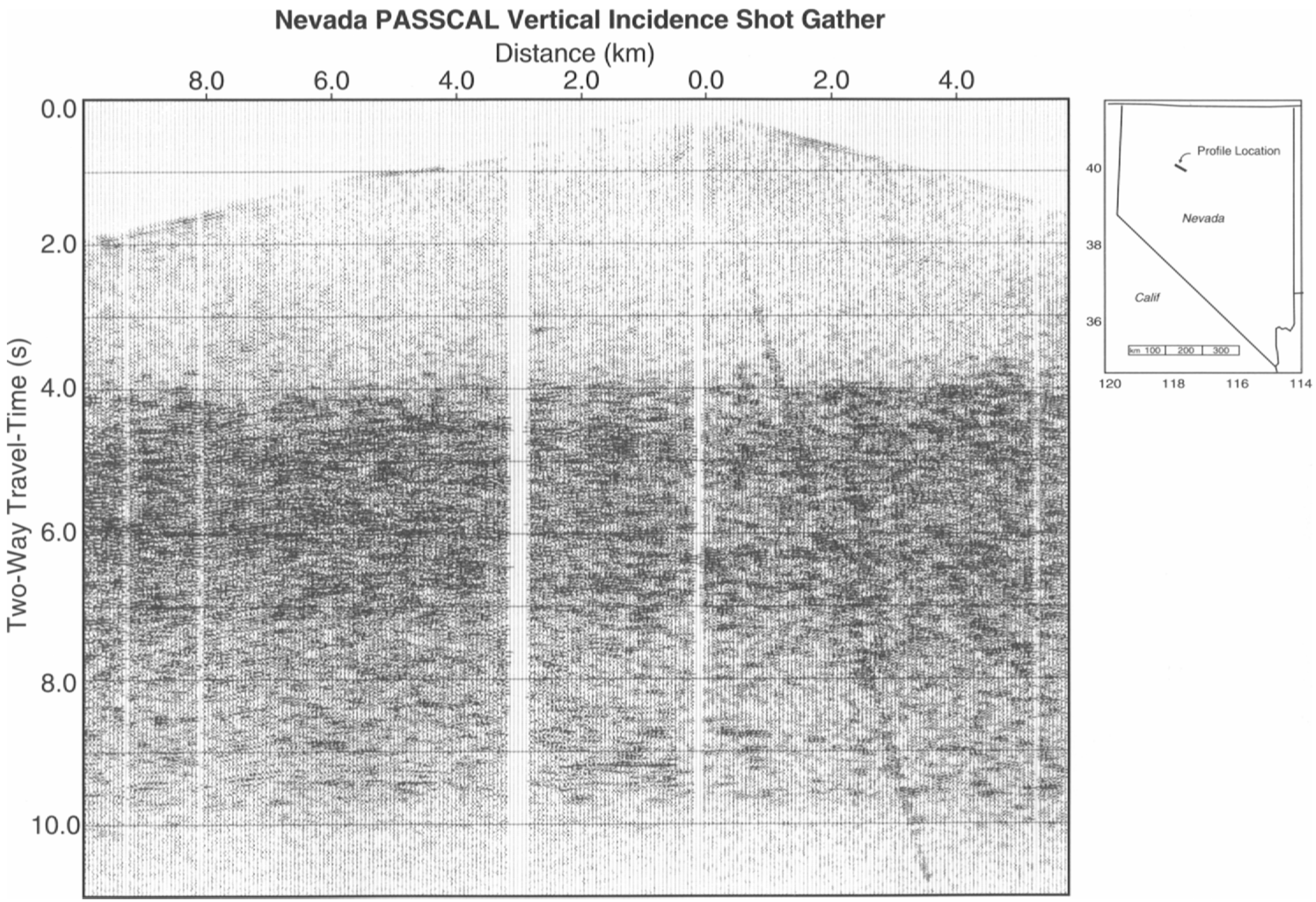

Fig. 7–7. Vertical incidence shot gather from the Nevada PASSCAL experiment in northern Nevada (after Jarchow et al., 1993). A relatively transparent upper crust lies above a highly reflective lower crust. The upper mantle is also transparent to vertical incidence seismic energy. This reflective texture is commonly observed in extended regions worldwide and is inferred to be a direct consequence of the extension process. These reflections were suggested to represent a combination of ductile shear and magmatic intrusions (Holbrook et al., 1991). A still molten horizontal intrusion was found at the Moho (Jarchow et al., 1993).

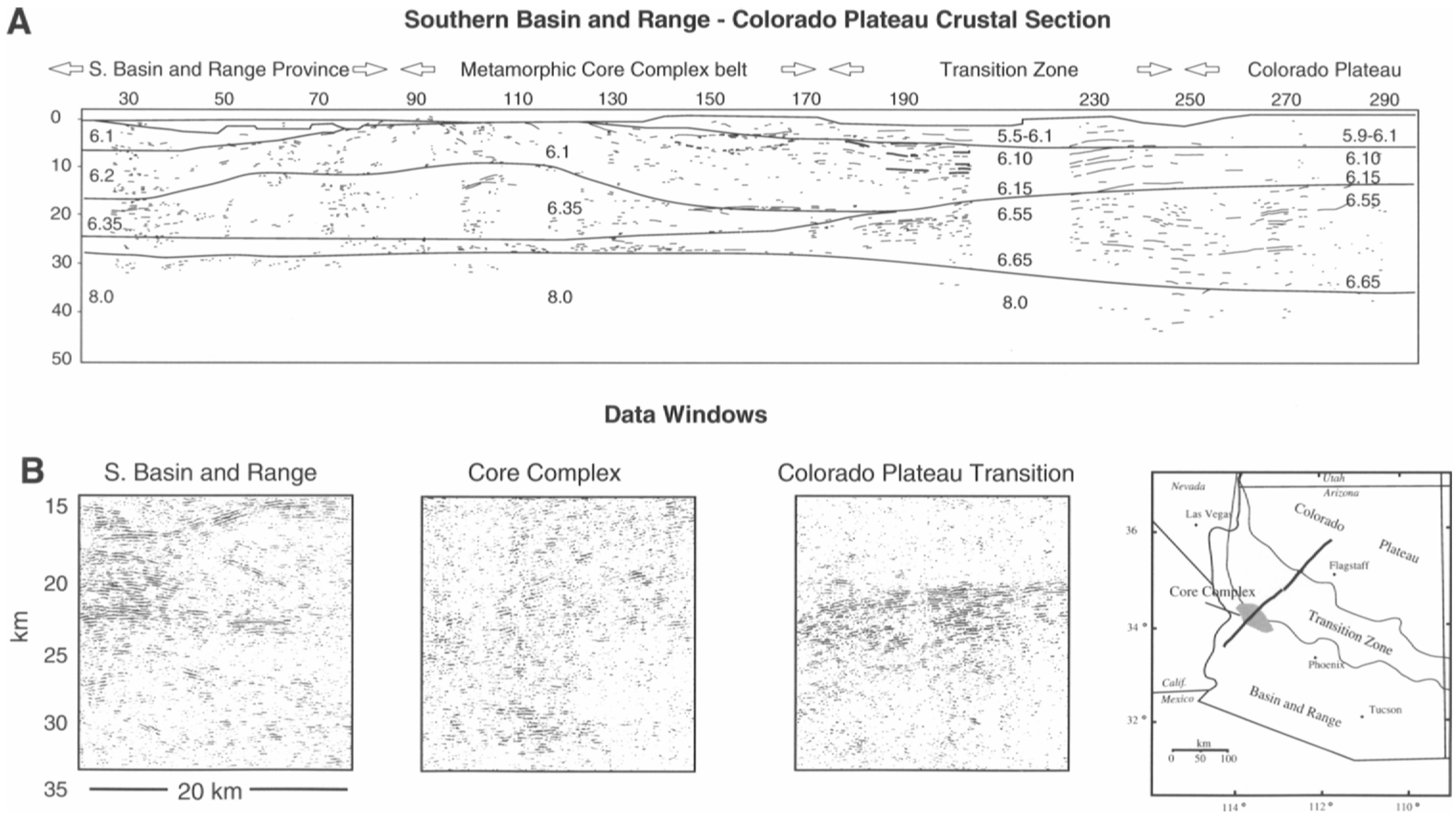

Fig. 7–8. Velocity model from a seismic refraction study in the southern Basin and Range that crossed "typical" Basin and Range crust as well as a metamorphic core complex (McCarthy et al., 1991) combined with line drawings of the COCORP Arizona seismic reflection profiles (after Hauser et al., 1987). Example data windows are shown below of the actual vertical incidence stacks. The southern Basin and range lower crust is highly reflective and is similar to the textures observed in the northern Basin and Range (Fig. 7–7). A thick welt of slow (6.35 km/s) lower crustal material underlies the Buckskin-Rawhide core complex, and reflection data from this area appears choppier and more disrupted than from beneath the typical Basin and Range or Colorado Plateau transition. Up to 4 km of mafic rock could have been added to the lower crust beneath the core complex if it was distributed in thin sheets (McCarthy and Parsons, 1994).

7.4.4.2 Seismic Refraction Studies: The Crustal Velocity Structure of the Basin and Range

Crustal seismic refraction studies in the Basin and Range province have been carried out along very similar transects to the seismic reflection profiles discussed above. Regional refraction profiles that cross Nevada east-west at 39° N, and north-south at about 116° W were carried out in the early 1960's and were reinterpreted by Prodehl (1979). These experiments had large station spacings, and did not provide much detail within the crust, but they were successful in determining the whole-crustal thickness. The wide-angle component of the 1986 Nevada PASSCAL experiment provides more crustal detail in northern Nevada (e.g., Catchings and Mooney, 1989; Benz et al., 1990; Holbrook, 1990). In the southern Basin and Range province, refraction studies were carried out above the Whipple Mountain metamorphic core complex in southeast California (Wilson et al., 1991), in central Arizona from the southern Basin and Range across the Colorado Plateau transition (Warren, 1969) and nearly coincident with the COCORP Arizona reflection profiles, northeast from the Colorado River into the Colorado Plateau (McCarthy et al., 1991).

There is some difference in the interpretations of the Nevada PASSCAL data (Catchings and Mooney, 1989; Holbrook, 1990; Benz et al., 1990), but the broad features are similar; the northern Basin and Range crust is thin, about 30 km thick, and does not vary more than 1–3 km along the profiles. A high velocity layer of varying extent was detected near the base of the crust in all interpretations, and magmatic underplating was suggested as the origin of those high-velocity rocks. Mid-crustal velocities were reported in the 6.0–6.2 km/s range, and have been suggested to be of granitic to granodioritic composition. Low velocity zones (~5.5 km/s) were found in the upper 10 km of crust, and were suggested to be fractured crust (Catchings, 1992) or silicic intrusions, underthrusted sediments, or high temperature zones (Holbrook, 1990). Reinterpretation of the early refraction data collected in the northern Basin and Range (Prodehl, 1979) ties well with the more recent profiles in terms of crustal thickness.

A fascinating result of wide-angle seismic profiling in the southern Basin and Range has been the crustal thickness determinations of the Whipple Mountain and Buckskin-Rawhide metamorphic core complexes in southeastern California (McCarthy et al., 1991; Wilson et al., 1991). Beneath the areas of the locally greatest extension, the crust maintains its thickness. A 10 to 15-km-thick welt of 6.35 km/s crust was identified on two crossing profiles just above a thin higher velocity (6.6 km/s) lower-crustal layer (Fig. 7–8). In addition, The upper-crustal low-velocity layer is thinner beneath the core complexes because mid-crustal rocks were exhumed on low-angle normal faults. The refraction studies of the Whipple Mountain and Buckskin-Rawhide metamorphic core complexes were the beginnings of the U.S. Geological Survey (USGS) Pacific to Arizona Crustal Experiment (PACE). The PACE profiles extend north of the core complexes across the Colorado Plateau transition zone, where the crust gradually thickens to a maximum of 39 km (Fig. 7–8). The zone of 6.35 km/s crust that was present beneath the core complexes is absent beneath the more weakly extended Colorado Plateau transition zone. A similar increasing crustal thickness across the Colorado Plateau transition zone was observed by Warren (1969) on a profile collected further to the east of the PACE transect. Braile et al. (1989), Mooney and Braile (1989), and Pakiser (1989) provide further discussion on the crustal velocity structure of the western United States. A discussion of the regional implications of crustal-structure variation can be found in Section 7.5.2.

7.4.4.3 Upper-Mantle—Velocity Structure from Earthquake-Source Studies

The upper mantle beneath the Basin and Range has been studied using a variety of techniques applied to long and intermediate period seismograms recorded by local, regional, and teleseismic earthquake and large explosive sources. Depending on the source range, receiver density, and phases analyzed, either an averaged one-dimensional velocity profile, or two- and three-dimensional variations in velocity structure are produced. A general result from studies of the Basin and Range province is an aver-

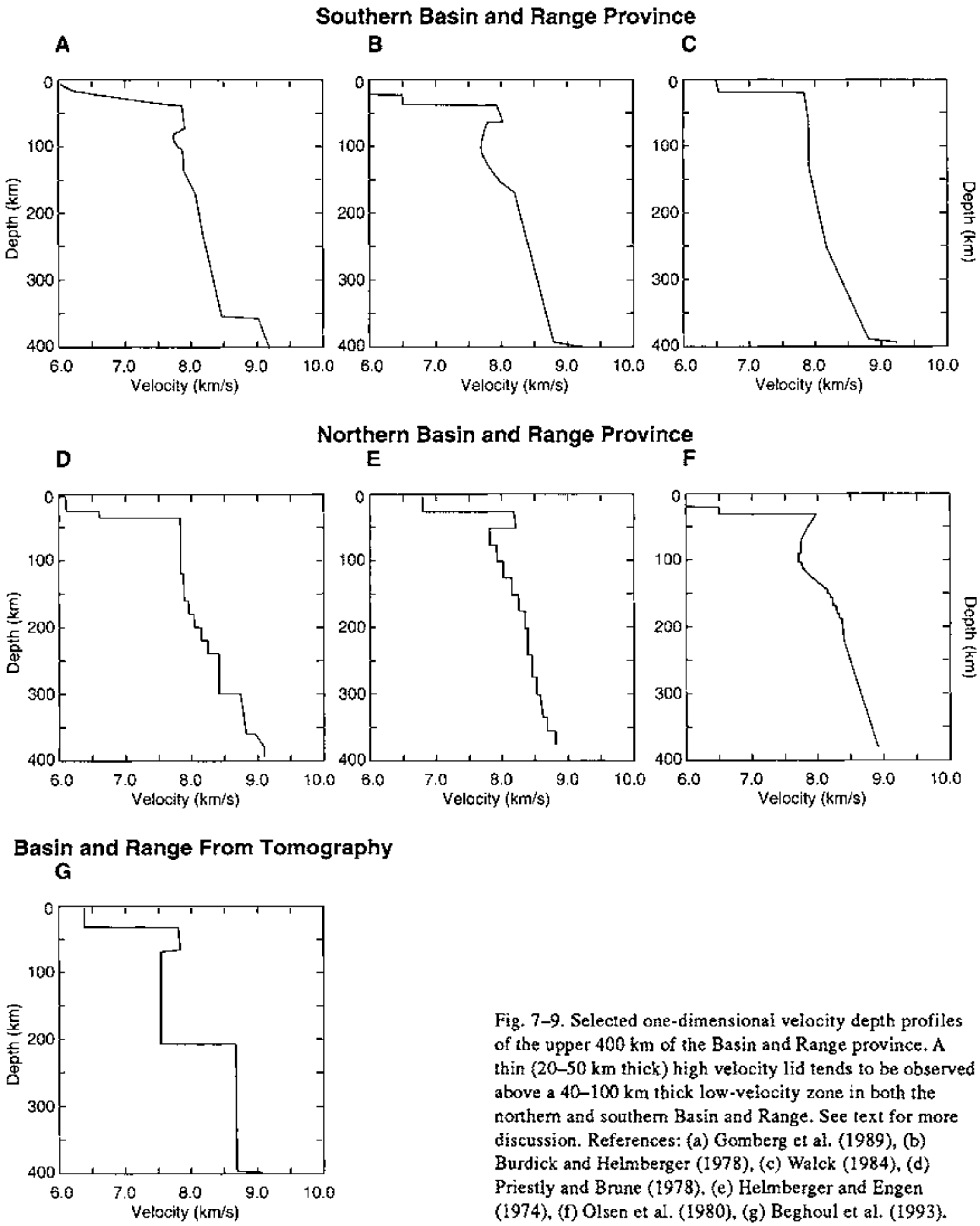

Fig. 7–9. Selected one-dimensional velocity depth profiles of the upper 400 km of the Basin and Range province. A thin (20–50 km thick) high velocity lid tends to be observed above a 40–100 km thick low-velocity zone in both the northern and southern Basin and Range. See text for more discussion. References: (a) Gomberg et al. (1989), (b) Burdick and Helmberger (1978), (c) Walck (1984), (d) Priestly and Brune (1978), (e) Helmberger and Engen (1974), (f) Olsen et al. (1980), (g) Beghoul et al. (1993).

age lower velocity upper mantle as compared with the interior continental craton, including a low velocity zone at or near the top of the upper mantle. The causes of these observations may include a thinner high velocity mantle lithosphere and a hotter, lower velocity asthenosphere than the continental interior. Variability in mantle structure within the Basin and Range province can be generalized to include the identification of the subducting Juan de Fuca slab beneath the northern part of the province and a low velocity asthenosphere connected with the Yellowstone hotspot. Teleseismic shear-wave arrivals in the Basin and Range often show significant splitting between the transverse and radial shear components, which may be the result of tectonic deformation in the mantle lithosphere and upper asthenosphere; thus splitting directions can help in the interpretation of mantle extensional dynamics.

One-dimensional Q or velocity-depth profiles for the Basin and Range crust and upper mantle generated from body and surface waves have been reported by Archambeau et al. (1969), Helmberger and Engen (1974), Burdick and Helmberger (1978), Priestly and Brune (1978), Olsen et al., (1980), Olsen and Braile (1981), Olsen et al., (1983), Walck (1984), Gomberg et al. (1989), and Al-Khatib and Mitchell (1991). The velocity profiles generally show about an 80-km-thick Basin and Range lithosphere overlying a thin low velocity zone, followed by a linear velocity increase with depth (Fig. 7–9). Upper mantle Q values tend to be lowest in the western Basin and Range, where the most recent magmatic and tectonic activity has occurred. One-dimensional shear-wave velocity models are also sensitive to recent volcanic activity, with the slowest low velocity zones recorded on the volcanic plateaus.

Two- and three-dimensional images of portions of the Basin and Range upper mantle have been created through tomographic inversion of direct mantle arrivals from regional sources (e.g., Hearn et al., 1991; Beghoul et al., 1993), and inversion of teleseismic residuals (e.g., Koizumi et al., 1973; Solomon and Butler, 1974; Iyer et al., 1977; Romanowiscz, 1979; Dueker and Humphreys, 1990; Biasi and Humphreys, 1992; Humphreys and Dueker, 1994). Tomographic studies show that the mantle lid beneath the Basin and Range is lower in velocity (< 7.9 km/s) than its margins (> 7.9 km/s). The mantle lid beneath the eastern Snake River Plain (track of the Yellowstone hotspot) is especially slow (< 7.6–7.8 km/s), presumably because of thermal softening of the upper mantle (Hearn et al., 1991). However, P_n velocities determined from controlled-source seismic studies tend to find faster velocities in the uppermost mantle at 8.0 km/s in the Basin and Range (e.g., Catchings and Mooney, 1989; Holbrook, 1990; McCarthy et al., 1991). This discrepancy may result from a shallower sampling by the controlled sources than from earthquake sources. Teleseismic residuals indicate a high velocity body deep beneath the northern and western Basin and Range province that is inferred to be the subducting Juan de Fuca plate or remnants of the Farallon slab (e.g., Koizumi et al., 1973; Solomon and Butler, 1974; Iyer et al., 1977; Romanowiscz, 1979; Biasi and Humphreys, 1992). The shallower velocity structure (< 250 km depth) identified through inversion of P-wave residuals shows northeast-trending low velocity zones beneath the eastern Snake River Plain and the St. George volcanic trend (chain of recent basaltic eruptions that crosses southern Nevada and Utah; Fig. 7–3c); the intervening zones of higher velocity upper mantle beneath central Nevada correspond to areas where magmatic activity peaked during Miocene time (Dueker and Humphreys, 1990; Biasi and Humphreys, 1992; Humphreys and Dueker, 1994). Upper mantle velocities as determined from a variety of methods appear to be slowest beneath areas that have seen tectonic and magmatic activity most recently, those areas near the eastern and western margins of the province.

It is generally agreed that upper mantle shear wave splitting results from tectonically imposed strain fabric in the mantle rocks (e.g., Ribe and Yu, 1991). Preferred alignment of olivine crystals causes the shear waves vibrating parallel to the elongate crystals to arrive faster than those vibrating perpendicular to the preferred-alignment direction. Thus the recorded fast direction is roughly parallel to the strain direction. The origin of the mantle strain is thought to be from internal continental deformation as well as from whole plate motion. Shear-wave splitting results are available from the northern Basin and Range (e.g., Savage and Silver, 1993) and the southern Basin and Range (e.g., Ruppert, 1992). Results from the northern Basin and Range show a fast direction of N 74° E with a 1 s difference between fast and slow arrivals (split) in the northwestern province, while in the northeastern Basin and Range the fast direction is about 45° different at N 72° W with a 1 s split (Savage and Silver, 1993). A similar variation in the horizontal stress direction from the northwestern to northeastern Basin and Range is observed (e.g., Zoback and Zoback, 1989), but the present-day stress directions do not parallel the fast or slow anisotropy directions. Savage and Silver (1993) suggest this mismatch may be because present day Basin and Range extension is too weak to overprint previous deformation fabric in the mantle. Two anisotropic layers (cumulative 0.7–0.8 s split) were postulated beneath the southern Basin and Range province in southern Arizona: a N 35° W fast direction in the 30–50 km depth interval (parallel to present-day extension) and an additional 80-km-thick zone extending into the asthenosphere with a fast direction oriented N 60° E, roughly parallel to North American plate motion (Ruppert, 1992).

7.4.5 Potential Field Studies

The northern Basin and Range province corresponds to a regional Bouguer gravity low, the boundary of which lies along the topographic edge between the northern and southern Basin and Range (Plate 7–2). The sources of the high topography and low Bouguer gravity in the northern Basin and Range both result from an upper-mantle low-density anomaly (e.g., Eaton et al., 1978). There is a bilateral symmetry in the long-wavelength gravity field of the northern Basin and Range province across a north-south center line (Plate 7–2). This increase in the Bouguer anomaly towards the Basin and Range margins was noted and compared with the outward migration of seismicity and magmatism into the margins by Eaton et al. (1978) and is suggested to result from asthenospheric upwelling beneath the central northern province. Isostatic residual maps have been created (e.g., Simpson et al., 1986; Jachens et al., 1989) that remove mantle effects from the Bouguer gravity so that crustal variations may be examined. These maps show evidence for low-density plutons at the heart of the bilaterally symmetric pattern in the Bouguer gravity map, possibly indicating a genetic link between anomalous upper-mantle and upper-crustal plutonism. Similarly, a large quantity of basalt intruded into the eastern Snake River Plain crust causes a strong positive crustal gravity anomaly in the northern Basin and Range.

Gravity data from the Basin and Range province have been used to address problems involving the extension process. Thompson and McCarthy (1990) pointed out that highly extended terranes such as metamorphic core complexes are not associated with large gravity anomalies, and the crust beneath them tends to be equally thick as less extended crust surrounding them. They suggested that at depth, rocks of equal density must have replaced the 10 km or so of upper crustal rock that was stripped off the top in order to prevent a strong gravity anomaly from being created. They further suggested that the low-density rocks must have been magmatically intruded, because inflow of ductile lower-crustal rocks would have emplaced enough dense material as to generate strong positive gravity anomalies which are not observed. However, Kruse et al. (1991) fit the Bouguer anomaly to a model of ductile lower-crustal flow from beneath the unextended Colorado Plateau to the extended Lake Mead region of the central Basin and Range. The Lake Mead region lies about 1 km lower than the Colorado Plateau and has a similar crustal thickness to the plateau, whereas many metamorphic core complexes are mountainous and are elevated above their surroundings. Thus examination of gravity data seems to indicate that lower-crustal flow and magmatism may operate independently or in concert to maintain crustal thickness during extension, depending on conditions within an individual terrane.

Aeromagnetic data from the Basin and Range provide information about crustal composition and evolution. A "quiet basement zone" extends from southern Nevada into Idaho (Mabey et al., 1978) and is centered along the line of symmetry identified on Bouguer gravity maps. The Snake River Plain stands out as a strong magnetic anomaly, particularly the western arm of the plain. The strength of the magnetic anomaly decreases along the eastern Snake

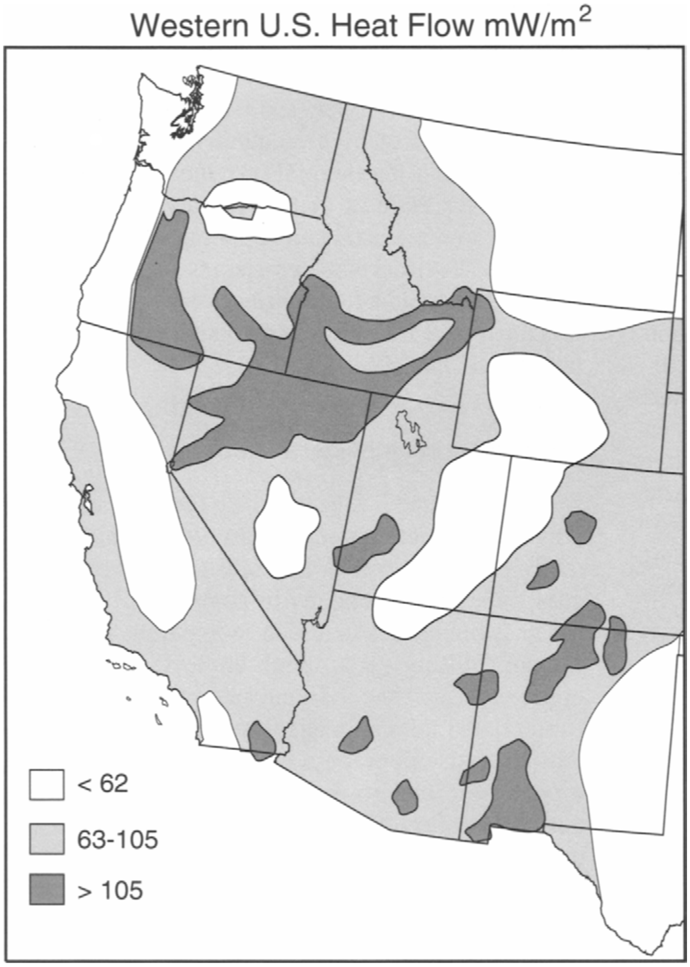

Fig. 7–10. Heat flow in the western United States after Morgan and Gosnold (1989). The darkest areas have the highest surface heat flow, and the lightest are the lowest.

River plain towards the Yellowstone caldera. Blakely and Jachens (1991) note that 46% of the surface area of Nevada has Mesozoic and Cenozoic igneous rocks in the upper 1 km of the crust. They also identify a roughly north-south trending 500-km-long magnetic anomaly across the middle of Nevada that has remained essentially linear since middle Miocene time. This feature has been termed the northern Nevada rift, and may represent a linear series of mafic dike intrusions (Zoback and Thompson, 1978; Zoback et al., 1994). The rift has been used as a marker for post-rift extension at an oblique angle to it (Zoback, 1978; Zoback and Thompson, 1978). The depth to the Curie temperature isotherm was estimated from magnetic anomaly properties in the state of Nevada, and the basal depth of the magnetic sources was observed to be shallowest in regions having the highest heat flow (Blakely, 1988).

7.4.6 Heat Flow and Magnetotelluric Studies

A result of recent tectonic extension and accompanying magmatism in the Basin and Range province has been an increase in surface heat flow relative to other more stable tectonic provinces in the western Cordillera such as the Colorado Plateau, Sierra Nevada, and Rocky Mountain foreland. Reduced heat flow in the Basin and Range is typically 50–100% higher than the surrounding stable provinces, and in places can be more than 300% greater (e.g., Battle Mountain high, northwestern Nevada) (Lachenbruch, 1978; Lachenbruch and Sass, 1978; Blackwell, 1978, Morgan and Gosnold, 1989). The relative highs and lows in reduced heat flow are shown in Fig. 7–10, though caution should be exercised when making regional interpretations of Basin and Range heat flow contours, as flow of ground water and variable magmatic advection can complicate matters considerably. The source of heat loss through the Basin and Range lithosphere has been suggested to be a direct result of elevation of hotter material closer to the surface caused by lithospheric thinning (e.g., Crough and Thompson, 1976; Lachenbruch and Sass, 1978; Artyushkov and Batsanin, 1984). However, constraints on the timing of extension, heat loss, and the observation that broad, unextended volcanic plateaus can be as hot as the Basin and Range province, leads to the conclusion that advective heat from magmatic sources must play an important combined role with lithospheric thinning in generating heat flow in the Basin and Range (e.g., Morgan and Gosnold, 1989; Mareschal and Bergantz, 1990). The southern Basin and Range province shows very similar heat flow values to the northern Basin and Range, though the metamorphic core complex belt along the Colorado River is relatively cold, while other regions of less pronounced Miocene extension remain hot, perhaps because 10 km of radiogenic upper-crust was stripped off the core complexes during extension (J. H. Sass et al., written communication, 1993).

Magnetotelluric studies in the southern Basin and Range province of southern California and Arizona find a highly conductive zone at depth that begins approximately at the Colorado River and continues to the east (e.g., Keller, 1989; Klein, 1990). A transect paralleling the USGS PACE seismic refraction and COCORP Arizona seismic reflection lines shows a deep zone of conductivity that may correlate with the top of the reflective lower crust and deepens beneath the Colorado Plateau transition (Klein, 1990). In the northern Basin and Range province, an intensive magnetotelluric survey was conducted across the Battle Mountain heat flow high, and high conductivities were observed through the crust at all depths, especially in Dixie Valley and the Carson Sink (Keller, 1989). A mid-crustal zone (~20 km deep) of high conductivity was also observed beneath the thermal province of the eastern Snake River Plain in Idaho (Stanley, 1982). A 70-km-wide lower crustal and upper mantle conductive zone was identified that corresponds with the roughly north-south trend of the northern Nevada Rift, and was attributed to diking and conductive permeable fractures (Chau, 1989).

7.5. Structure and Interpretation

The Basin and Range province presents some key crustal and mantle structural problems that bear directly on the extension and uplift processes. I discuss three such problems in some detail here: (1) the possible causes that lead to apparent low-angle normal faulting versus high-angle faulting in the upper crust, (2) the maintenance of relatively uniform crustal thickness despite strongly varying extension and topography, and (3) the variations in topographic elevation between the southern and northern Basin and Range province despite the nearly uniform crustal thickness and average magnitude of extension in the two sub-provinces.

7.5.1 Low-Angle vs. High-Angle Normal Faulting

Anderson's (1951) theory provided a general framework to describe faulting in relation to the ambient stress field in the Earth's crust. The theory predicts that when the vertical lithostatic load is the greatest principal stress, normal faulting ensues at an angle ~45° to 70° from vertical, when the difference between the horizontal least principal stress and vertical greatest principal stress exceeds the shear strength of the rocks. As more faults are investigated world-wide, it is increasingly clear that Anderson theory alone cannot adequately describe many observed faults. For example, shallow dipping to horizontal fault planes are commonly observed, often with extreme normal displacements. Because these faults are shear failures that respond to the local stress field, apparently either the greatest principal stress direction deviates from the vertical (e.g., Bartley and Glazner, 1985; Bradshaw and Zoback, 1988; Melosh, 1990), or a steeply dipping fault plane rotates to a more shallow dip after displacement (e.g., Davis, 1983). Reactivation along ancient low-angle fault planes is a less likely explanation because in most cases the shear strength of a plane of weakness improperly oriented to the principal stress axes exceeds that for a new fault in fresh rock along a more favored plane (Sibson, 1985). There is clear evidence that rotation followed by initiation of new fault planes occurs, as in the case at Yerington, Nevada (Proffett, 1977). However, evidence from structural reconstructions indicates that many detachment faults begin and propagate at low angles, including the Whipple Mountains (Yin and Dunn, 1992) and Chemehuevi Mountains of southern California (Miller and John, 1988), the Harcuvar Mountains of central Arizona (Reynolds and Spencer, 1985), and in the Mormon Mountains of Nevada (Wernicke et al., 1985).

The expected pattern for brittle extension with a vertical greatest principal stress is finite motion along steeply dipping fault planes, with a new plane forming when it is no longer efficient to continue motion along the first (much like the middle Miocene episode of block faulting in the Basin and Range province). Historical earthquakes have focal mechanisms that indicate steeply dipping fault planes (>40°) (e.g., Jackson, 1987). When lower angled faults occur, they tend to expose sharp divisions between brittle deformation in the upper plate and ductile deformation in the lower plate; this has invited the suggestion that the low-angle faults represent the brittle-ductile transition (e.g., Gans et al., 1985). Models involving isostatic uplift ("rolling hinge" models) suggest that unloading caused by movement on a normal fault causes upwarp of the footwall, which rotates the initially steep fault plane towards the horizontal (e.g., Heiskanen and Vening Meinesz, 1958; Spencer, 1984; Wernicke and Axen, 1988). Observational evidence suggests that isostatic rebound of local features such as the footwall of a normal fault occurs in the middle crust, rather than involving the whole crust. For example, analysis of gravity data suggests that the load differences between the relatively narrow (~20 km wide) basins and ranges in the western Cordillera of the United States are supported by the strength of the crust (e.g., Eaton et al., 1978; Kruse et al., 1991), which implies that isostatic compensation has occurred by midcrustal shear flow, rather than in the mantle. Seismic reflection profiles from the northern Basin and Range province tend to support that result, showing a flat Moho beneath both basins and ranges (e.g., Klemperer et al., 1986). Footwall rebound should occur equally on every steep normal fault of approximately the same offset, which indicates that the observed variation in dip-angle is likely a consequence of heterogeneities in composition, rheology, or stress distribution.

The stress field and rheology of the upper crust may be influenced towards conditions favoring low-angle normal faulting by magmatism (e.g., Lister and Baldwin, 1993; Parsons and Thompson, 1993). Advected heat from intruded magma could raise the brittle-ductile transition temporarily, and shear along the ductile zone (e.g., Melosh, 1990) or shear stresses imposed directly by the intrusion (Parsons and Thompson, 1993) might cause a rotation in the stress field favoring low-angle faulting. Any model for in-situ low-angle faulting must include some source of shear that drives the faulting, either gravitational as in the case of Gulf of Mexico faults (e.g., Bradshaw and Zoback, 1988), or some other source where topographic inclination is not likely to be a factor, such as in the metamorphic core complexes in the Basin and Range. Furthermore, models for metamorphic core complex development must be consistent with the following observations: (1) the crust beneath core complexes is often as thick as surrounding less extended terranes, (2) core complexes lack strong Bouguer gravity anomalies, thus regardless of its source, the material that maintains crustal thickness beneath the core complexes must be on average the same density as the whole-crustal average, and (3) the exposed core rocks are typically warped upwards into mountainous antiformal structures rather than buried beneath thick sedimentary basins.

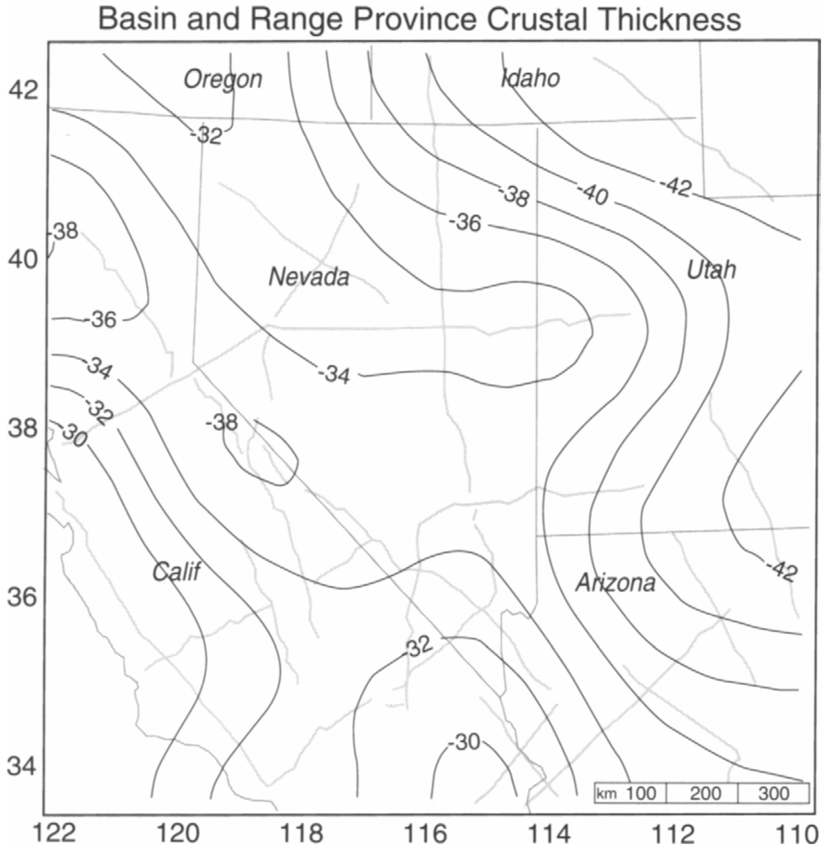

Fig. 7–11. Contour plot of the crustal thickness of the western Cordillera where estimates from seismic refraction profiles are available (Fig. 7–6) (shown as gray lines; Warren, 1969; Prodehl, 1979; Catchings and Mooney, 1989; McCarthy et al., 1991). This plot is similar to compilations by Pakiser (1989) and Jones et al. (1992).

7.5.2 Basin and Range Crustal Extension and Thickness

As the number of seismic refraction studies in the Basin and Range increased, it became clear that the crust within the extended province is remarkably uniform despite strong variation in the timing and magnitude of stretching from terrane to trerrane (Fig. 7–11). Either the pre-Tertiary Basin and Range crust was a minimum of 60 km thick, and up to 120 km thick in places, or crustal material was added during extension. Long-term inter-plate convergence and subduction very likely did increase the crustal thickness of western North America during the Paleozoic and Mesozoic Eras (e.g., Bird, 1984), but it would have to have exceeded the present thicknesses of the Rocky Mountains and Colorado Plateau if the present Basin and Range crustal thickness evolved through closed-system pure shear of the Cordilleran crust. Gans (1987) inferred a 40 to 50 km pre-extension crustal thickness for the northeastern Basin and Range province, and reconciled the estimated 77% surface extension with the present 30–35 km thickness by proposing an open-system mode of extension that included lower-crustal ductile flow and the addition of a 5-km-thick magmatic layer.

The maximum focal depth of earthquakes in the Basin and Range province is about 15 km, an indication that the crustal layer beneath the brittle upper crust responds to regional extensional stresses by aseismic creep, or ductile flow. The high observed surface heat flow indicates that the ductile layer may have increased mobility through heat conducted from the mantle and/or advected by intruding magma. Buck (1991) suggested that lithospheric rheology controls extensional modes. When thermal weakening dominates and lower crustal flow occurs very fast, then core complexes form; otherwise either a wide rift occurs if the lithosphere is moderately hot and weak, or a narrow rift forms if the lithosphere is cold and strong. The role that the ductile layer may play in maintaining crustal thickness during extension is not well constrained, nor is the amount of ductile lower-crustal material that may have been added to the extending Basin and Range crust from more stable unextended terranes such as the Colorado Plateau or the Sierra Nevada. Perhaps some indication may come from the patterns of seismic reflectivity that are commonly observed in the lower crust of extended terranes worldwide. As was shown in Section 7.4.4.1, the lower crust of the Basin and Range province can be highly reflective at vertical incidence (e.g., Figs. 7–7, 7–8). The reflective lower crust does not extend very far into the more stable terranes along the Basin and Range margins such as the Sierra Nevada, the northern and southern Colorado Plateau, or the Mojave block, all regions where pre-Tertiary structures seem to be preserved (e.g., Cheadle et al., 1986; Allmendinger et al., 1987; Howie et al. 1991). If the lower-crustal reflectivity is the result of shearing in the lower crust, and significant flow from stable blocks occurred, than such fabric would be expected beneath those stable blocks. The absence of reflectivity from the bordering stable terranes suggests that significant flow from these regions has not occurred. A further complication in interpreting lower-crustal reflectivity is the strong possibility that at least some of the reflectors are caused by Tertiary magmatic intrusions into the extended crust.

The Moho is an often-suggested locus of magmatic underplating (e.g., Furlong and Fountain, 1986; Matthews and Cheadle, 1986; Bohlen and Mezger, 1989; Bergantz, 1989; Wilshire, 1990). The seismic velocities from the highly reflective lower crust of the northern Basin and Range province are high (7.1–7.3 km/s) in places, and are lower (~6.6 km/s) in other parts, perhaps indicating variable input of mafic magma into the lower crust there (e.g., Holbrook et al., 1991). Lower-crustal velocities tend to be slow in the southern Basin and Range province, though velocity constraints allow up to a 4 km thickness of mafic intrusive rocks if it is distributed within the lower-crustal layer in thin sheets (McCarthy and Parsons, 1994). While underplating or intrusion of distributed horizontal mafic sheets is often suggested as a means to maintain crustal thickness and advect heat into the lower crust to stimulate flow there, little attention is given to how such sheets might be emplaced in an extensional environment that should favor vertical rather than horizontal intrusions. The rheologic boundaries of the lithosphere such as the Moho may be traps where basaltic magma is more likely to orient horizontally because the rheologic differences lead to long term changes in the stress magnitude across them. While the upper-most mantle lid is likely more rigid than the lower crust it is seldom seismogenic in extending regimes. The commonly observed association of extension with magmatism and high heat flow is perhaps an indication that dike intrusion along with some amount of ductile deformation is the means by which upper-mantle extensional strain occurs. The ductile lower crust develops less deviatoric stress and may become over-inflated by dike intrusion as compared with the stronger upper mantle (e.g., Parsons et al., 1992b) (Figure 7–12). Subsequent dikes in the mantle reach the ductile lower crust in which the horizontal stress exceeds the lithostatic load, and the magma is hence forced to intrude horizontally, underplating the Moho. Multiple occurrences of this cycle may be important in the evolution of the lower crust and Moho in extending terranes.

Models of whole-lithospheric extensional flow identify two wavelengths of necking instabilities, a local one at the scale of basin-range spacings that is accommodated by middle and lower crustal flow, and a province-wide thinning of the lithospheric mantle (e.g. Froidevaux, 1986; Zuber et al., 1986). That

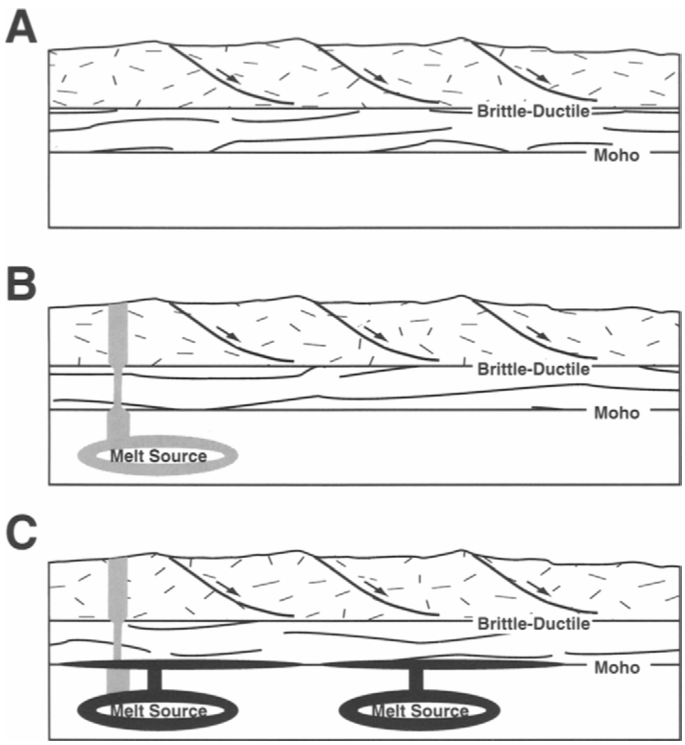

Fig. 7–12. Sequence of events leading to horizontal intrusion and underplating at the crust-mantle boundary. (a) In an extending regime the deviatoric stress (difference of maximum vertical and least horizontal) accumulates in the more brittle upper crust and upper mantle. The more ductile lower crust tends to flow and relax the deviatoric stress before much can accumulate there.(b) A vertical dike initiates, and thins as it crosses the Moho, because of the smaller deviatoric stress. The strain caused by the dike in this case imposes a stress which exceeds the residual deviatoric stress in the lower crust. The dike continues into the brittle upper crust where it widens, and combines with normal faulting to accommodate the deviatoric stress there. (c) A second pulse of dike intrusions form in the mantle and propagate upwards, but the dikes encounter a compressive regime in the lower crust. The melt is induced to intrude horizontally at the crust-mantle boundary because of the new stress conditions there. The second set of dikes accommodates the remaining deviatoric stress in the mantle, which may begin accumulating again with continued extension and the whole cycle may repeat many times. Multiple cycles of this process would cause the lower crust to become more mafic, and would help extending crust to maintain its thickness as well as influence the character of the Moho.

predicted local variability of extensional flow combined with the observed seismic velocity and reflectivity observations can perhaps lead to a unified conclusion. Since it seems that significant flow from outside the Basin and Range province is precluded by the lack of strong patterns of reflectivity in its margins, then the only sources that could have maintained Basin and Range crustal thickness are a thick pre-extension crust, and/or intrusion of magma from the mantle. Furthermore, the variable seismic velocities in the lower crust of the Basin and Range indicate that underplating probably occurs as distributed thin sheets rather than as a thick, contiguous layer. The scale length of lower-crustal reflections from beneath the Basin and Range is typically short (<5 km), perhaps an indication that ductile crustal flow has acted to disrupt the intrusive sheets into small lenses, while also evening out crustal thickness locally, pulling rock from beneath the ranges into the crust beneath the basins. Underplating as thin sheets is more effective in conducting heat quickly into the lower crust than would be a massive underplated layer. Thus the granulite xenoliths observed from beneath the southern Basin and Range may have been heated by a relatively smaller volume of mafic melt than was proposed by Hayob et al. (1989), which is more consistent with the low observed $^3He/^4He$ ratios (e.g., Torgersen, 1993).

7.5.3 Isostatic Constraints on the Basin and Range Upper Mantle Structure

The mantle beneath the Basin and Range province can be investigated if the thickness and density of the crust are known. If a good estimate of these two parameters can be made, then it is possible to estimate the crustal contribution to uplift and thereby isolate the remaining mantle contribution. There is a strong indication that the southern and northern Basin and Range mantle are behaving quite differently, since the crust in the two regions is similar in thickness and average velocity, while their topographic elevations differ by about a km. If the crustal buoyancy alone cannot float the lithosphere enough to account for the difference in elevations, then a mantle contribution to the northern Basin and Range elevation is indicated. The crustal-thickness information for the western United States from seismic refraction surveys (Fig. 7–11) was combined with information on upper-crustal density variation from a basin-stripped isostatic residual gravity map (Saltus, 1991). These gravity data give an indication of the relative density variation of the upper-crustal basement rocks because the effects of sedimentary basins and regional trends have been removed. Density perturbations of the upper-crustal rocks can be approximated with the relation $\rho \approx (g/0.04L_{uc})$ (e.g., Simpson and Jachens, 1989), where g is gravity in mGal, and L_{uc} is an assumed upper-crustal thickness (15 km). Lower-crustal densities are not delimited by the data and were assigned a uniform value of 2.9 g/cm^3; the lower-crustal layer is too thin for even strong density variations within it to cause the large observed mass deficit. Combining the crustal-thickness and upper-crustal density data results in the approximate crustal contribution to uplift across the western Cordillera; this when compared with the topography, indicates the mantle contribution to uplift (Fig. 7–13). Regional isostasy is assumed because the free air gravity anomaly averages zero across the region (e.g., Thompson and Zoback, 1979).

The upper-mantle mass per unit area is found by dividing the study area into a grid of isostatic columns. These columns are calculated using a two-layer isostatic expression with a known crustal layer, and a second mantle layer that extends down to an assumed local iso-density asthenosphere at a depth of ~200 km. The problem can be viewed as a group of 200-km-deep columns each containing the unknown boundary between the lithosphere (of unknown density) and the fixed-density asthenosphere. Since the crust is treated as a known quantity, and the topographic elevation is known, then the mass contained within the lithospheric mantle part of the column required to float it to its individual height can be calculated. The mantle mass per unit area of a given column (ΔM_n) is found by comparing it to a reference column, which is the column in the study area that contains the greatest amount of mass (and hence has the lowest mantle buoyancy), as

$$\Delta M_n = \left(I_n - I_{rd}\right) ,$$

where

$$I_n = \rho_a E_n - \left(\rho_a - \rho_c\right) L_{cn}$$

is a modified form of the isostatic equation (e.g., Lachenbruch and Morgan, 1990). The variable L_{cn} is the thickness of the crust in column n, ρ is density with the a and c subscripts denoting asthenosphere and crust respectively (ρ_a = 3.2 g/cm^3; variation in the fixed asthenosphere density does not affect relative difference between columns), and E_n is topographic elevation .

On the basis of isostatic calculations, Figure 7–13 shows that a broad low-density anomaly underlies the northern Basin and Range province, extending from the Sierra Nevada in the west to the Colorado Plateau in the east. Similar results were achieved by Jones et al. (1992), who used a slightly different method. The peak of the anomaly corresponds approximately to the center of the northern Basin and Range province, with the bulk of the anomaly to the south beneath the central Basin and

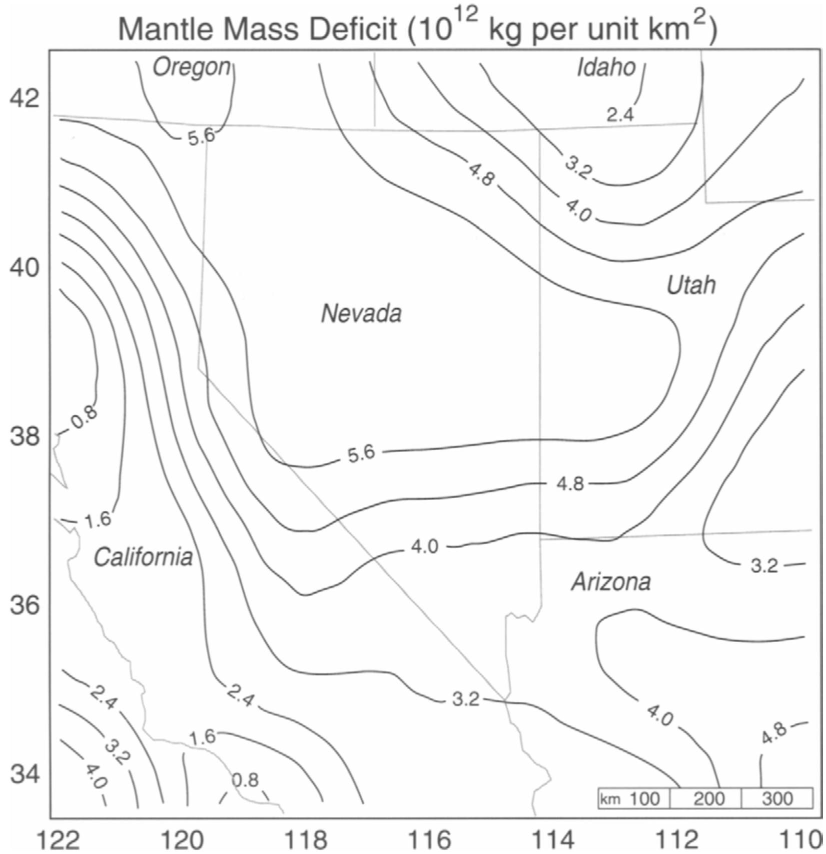

Fig. 7–13. Contour plot of relative variation in upper-mantle mass beneath investigated part of the western Cordillera. Values were calculated using a modified form of the isostatic equation with crustal depth information from seismic-refraction profiles (Fig. 7–11). The free-air gravity anomaly across the region is near zero, which implies that the area is in isostatic equilibrium. The thin crust of the northern Basin and Range province associated with its high elevation requires an underlying low-density upper mantle if isostasy is to be satisfied. The similar crust of the southern Basin and Range is a km lower in elevation, a possible indication that the mantle asthenosphere is different from north to south.

Range. The inferred Yellowstone plume track (eastern Snake River Plain) does not appear as a low-density anomaly on this image because the crust there has been largely replaced by more dense basalt (e.g., Sparlin et al., 1982), which has caused subsidence. The sum of the mantle isostatic mass deficit is $\sim 3.6 \times 10^{18}$ kg beneath the region investigated. Extension and thinning of the Basin and Range lithosphere allowed the asthenosphere to rise and take its place, causing some of the mass deficit. Earthquake-source methods that measure the variation in lithospheric thickness between the Basin and Range province and the unextended Colorado Plateau converge on a difference in the mantle-lid thickness that ranges between 15 and 50 km across the two provinces (see Section 7.4.4.3 or Iyer and Hitchcock, 1989; Beghoul et al., 1993). If a reasonable density contrast between asthenosphere and lithosphere of 0.05 g/cm^3 is used (e.g., Thompson and Zoback, 1979; Lachenbruch and Morgan, 1990) then the mass deficit caused by lithospheric thinning amounts to $\sim 4.8 \times 10^{17}$–1.6×10^{18} kg, or 13%–44% of the total anomaly. If the entire anomaly is attributed to lithospheric thinning over an adiabatic asthenosphere, then a mantle lithospheric layer ~112 km thick would need to have been removed from the Basin and Range province across the study area to account for the entire estimated mass deficit.

The mantle lid as measured in the northern and southern Basin and Range province is very similar (Fig. 7–9), which implies that some other cause of low density mantle besides mantle lithospheric thinning has occurred in the northern Basin and Range. Saltus and Thompson (1993) investigated the topographic difference between the northern and southern Basin and Range using gravity data and found that only half of the isostatic support could possibly come from the crust, while thermal expansion and phase changes were unlikely sources of the density anomaly in the mantle. Instead, they suggested that the Yellowstone plume may be responsible for the low density northern Basin and Range mantle. The possible role of the Yellowstone plume in Basin and Range extension is discussed further in Section 7.6.4.

7.6. Tectonic Evolution: Sources of Basin and Range Extensional Stress

The extensional stresses that caused the opening of the Basin and Range province are the consequences of pre-Tertiary and Tertiary tectonic events. Possibly both passive and active rifting have occurred at various stages of Basin and Range extension. As was discussed briefly in Section 7.1.1, tectonic cycles have worked and reworked the western Cordilleran lithosphere since it was a passive margin in the Proterozoic Eon. It would be difficult to quantify exactly how much the lithosphere was weakened relative to the cratonic interior by those tectonic cycles, but they have very likely defined the breadth of Basin and Range extension, since broad rifting is predicted in weak lithosphere, while narrow, more typical rifting is predicted in cold, strong lithosphere (e.g., Buck, 1991). It is a common coincidence to observe extensional deformation overprinting earlier compressional structure throughout the Basin and Range province (e.g., Coney, 1987; Wernicke, 1992). Many of the sources of extensional stress can be attributed to the varying configurations between the North American and the Pacific, and the now-extinct Farallon, oceanic plates.

7.6.1 North America – Farallon Plate : Back- and Intra-Arc Extension

At about 43 Ma a major global plate reorganization caused a change in the Pacific plate motion relative to North America (e.g., Engebretson et al., 1985). The Farallon plate still intervened between North America and the Pacific plate at this time (Fig. 7–3), and was isolated from the global system by a spreading center between it and the Pacific plate and a subduction zone along the North American plate boundary. The motion of the Farallon plate may also have begun to slow independently of the global system because the Pacific-Farallon spreading ridge drifted east relative to North America, causing the subduction of increasingly younger oceanic lithosphere until a plate with positive buoyancy was forced beneath North America (e.g., Engebretson et al., 1984). The rate of Farallon-North American convergence at 40° N was still at its peak at 50 Ma, but began a rapid decline shortly after that (e.g., Engebretson et al., 1984). Heaton and Kanamori (1984) noted a worldwide tendency for back-arc basins to form behind subduction zones with rapid convergence and low coupling. Thus the suggestion that the early stages of northern Basin and Range province extension that began in southern Canada and the northwestern United States resulted from back-arc extensional stresses (e.g., Zoback et al., 1981) seems valid. Stratigraphic constraints and trace-element signatures of volcanic rocks that erupted at about 30 Ma in northern Mexico indicate that the initial stages of extension there occurred in a back- or intra-arc setting (Cameron et al., 1989). When the Farallon-North American convergence rate slowed because the oceanic slab buoyancy increased, that style of extension seems to have died out in the northern Basin and Range province. The ignimbrite volcanic event in central Nevada, the closing of the Laramide magmatic gap, and associated extension correlates in time with the onset in slowing of the convergence rate between the Farallon and North American plates and a possible change in the angle of subduction. Speculatively, this stage of extension may have been a brief period of active rifting, driven by a sudden burst of magmatism associated with renewed asthenospheric contact beneath Nevada and

Utah after the slab angle steepened. Elston (1984) suggested that an "extensional orogeny" caused the ignimbrite event, and that the subduction of hot, young oceanic lithosphere enhanced the tendency of continental lithosphere to extend towards its margin through thermal input.

7.6.2 Extensional Spreading Resulting from Orogenic Over-thickening

A consequence of unstable over-thickening of continental lithosphere is that it will attempt to spread back out to its pre-compressional thickness. The over-thickened lithosphere can be thought of as storage of potential energy poised to be released. In order for that release to occur, the outer boundaries of the thickened lithosphere must become less confined, to allow room for spreading, or the topographic head must overcome those constraints. Widespread compressional features in the western Cordillera indicate that lithospheric thickening occurred during long-term subduction and reached far inland during the Laramide orogeny, perhaps because of low-angle subduction of the Farallon slab (e.g., Dickinson and Snyder, 1978; Bird, 1984). Coney and Harms (1984) reassembled the Tertiary extension of the Basin and Range assuming extension on the order of 40–60%, and found the pre-extension thickness to be about 40–60 km, similar to thicknesses beneath the Colorado Plateau and Rocky Mountains. Further explanation is required to explain why the over-thickened Basin and Range province lithosphere should have extended so readily while the equally thickened Colorado Plateau and Rocky Mountains have not. Coney (1987) and Sonder et al. (1987) suggested that a softer lithospheric rheology caused by the ignimbrite event may have stimulated lithospheric outflow and allowed preferential spreading in the Basin and Range; they further suggested that this spreading may have been enhanced by decreased compressional stresses applied at the continental margin because of slowing Farallon subduction. Harry et al. (1993) proposed through finite-element modeling that the orogenic over-thickening may have caused upper mantle weakening that focused extensional strain and re-equilibration of lithospheric thickness in the Basin and Range province. They also suggested that the marginal highlands of the Sierra Nevada and Colorado Plateau were outside the regions of upper mantle weakening and hence were not extended.

7.6.3 Basin and Range Rifting as a Result of Pacific - North American Plate Divergence

The global-circuit and hotspot-reference plate reconstruction models (Engebretson et al., 1985; Stock and Molnar, 1988) show that the Pacific plate has drifted to the northwest relative to a reference fixed on the interior of North America since 42 Ma. The rate of longitudinal displacement (the westerly component) has been about 30 km/m.y., depending on the latitude examined. The half spreading rate on the Pacific-Farallon ridge was slightly greater over the same time period, allowing the ridge axis to migrate slowly eastward towards North America. The initial contact of the eastern accreting edge of the Pacific plate with the western edge of the North American plate occurred at 29 Ma just south of the Pioneer fracture zone (Severinghaus and Atwater, 1990). Since then the Pacific plate has moved west by about 10.2° of longitude, which corresponds to slightly over 900 km at the latitude of 38° N. This compares to 8.3° of latitudinal drift during the same time period. The segment of the North American edge at the initial contact point has kept pace with the rapid west drift of the Pacific plate for the last 29 m.y. The Pacific-North American plate boundary evolved in the last 29 Ma from a short simple contact at the subduction interface to the present complex zone, over 2,300 km long, that partly resides within the continent and partly includes a small new ocean in the Gulf of California (Atwater, 1970; Lonsdale, 1991). Atwater (1970) explained the plate boundary evolution in the context of the migrating-triple-junction paradigm. However, Lonsdale (1991) has convincingly shown that long, extinct ridge segments and the corresponding fragments of the Farallon plate are still preserved offshore along most of the Pacific-North American plate boundary, limiting the applicability of the migrating triple junction hypothesis to the Californian system and indicating that the boundary lengthened primarily through a series of plate-capture events. With each

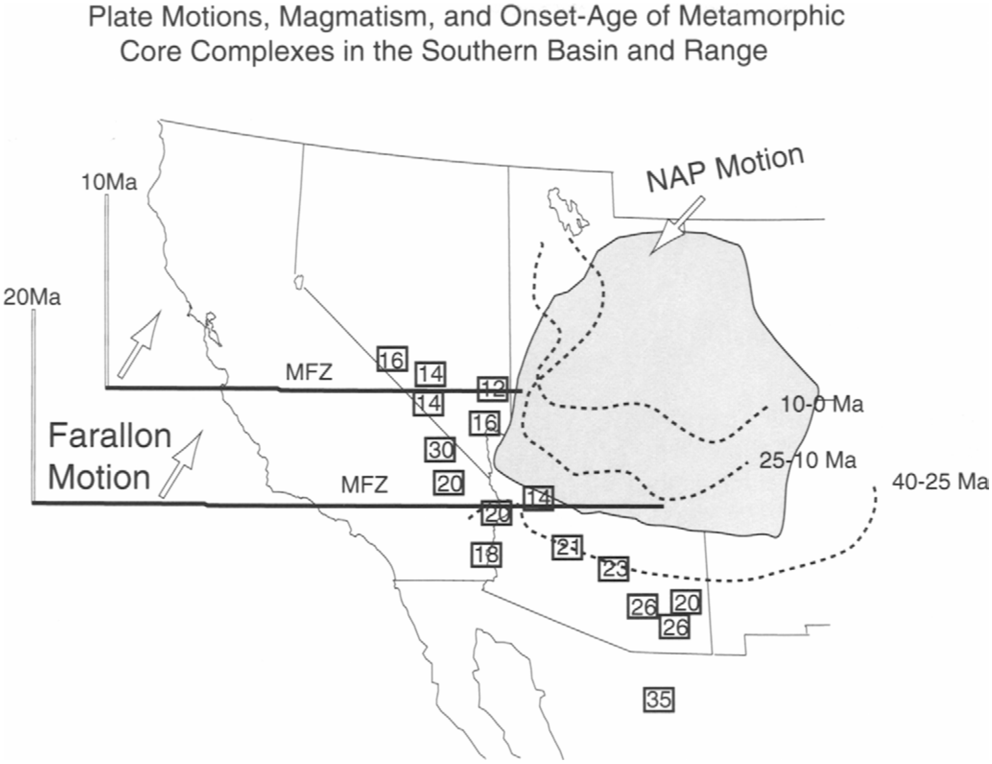

Fig. 7–14. Predicted locations of the coherent Farallon slab shown with the locations and timing of onset of core complex extension and magmatism. The dashed lines mark the local northward extent of magmatic activity with time after Armstrong and Ward (1991), and the boxed numbers are approximate time since core-complex initiation (in Ma) as summarized by Axen et al. (1993). The gray shaded area is the uplifted Colorado Plateau. These tectonic events are part of a broad northwest-younging trend that parallels the continental margin, and seems related to the change in subduction style.

event, a longer segment of the North American continent and the corresponding subducted fragment of the Farallon plate were exposed to the rapid west drift of the Pacific plate. Cordilleran North America expanded westward to fill the burgeoning gap between the rigid Pacific plate and the cratonic interior of North America (e.g., Bohannon and Parsons, submitted, 1994).

Many researchers have noted an apparent tie between the onset of extension in the southern Basin and Range province and the termination of Farallon plate subduction beginning at about 30 Ma (e.g., Atwater, 1970; Engebretson et al., 1984; Glazner and Bartley, 1984; Severinghaus and Atwater, 1990; Ward, 1991; Axen et al., 1993). Axen et al. (1993) and Gans et al. (1989) show a space-time migration of activity in this belt that closely corresponds to the northward progression of the Pacific plate contact zone between 15 and 30 Ma. It began at approximately 30 Ma at precisely the latitude of the initial contact. The space-time migration of the initiation of extension follows the position of the northern limit of the plate contact, and the migration of cessation corresponds well with the northern limit of active spreading taking place south of the Farallon fracture zone (Figure 7–14). By 15 Ma the Pacific-North American contact zone had lengthened to include all of the continental edge from the southern borderland rift of Crouch and Suppe (1993) to the latitude of southern Nevada. Most of the extension that was active in the Basin and Range province up to 15 Ma was taking place somewhere directly inland of the contact zone (Axen et al., 1993). By 10 Ma that pattern had probably changed somewhat because a large amount of regional extension was taking place in the Great Basin that was north of the inland projection of the Mendocino fracture zone (e.g., Hamilton, 1987; Gans et al., 1989; Wernicke, 1992). The direction of least principal stress (and extensional strain) had rotated about 45° by this time to a more east-west direction (Zoback et al., 1981).

7.6.4 Role of Yellowstone Plume in Basin and Range Extension?

During the past 16–17 m.y., the North American plate has moved southwest over the Yellowstone plume, leaving the Snake River Plain behind as its track (Morgan, 1972; Armstrong et al., 1975, Pierce and Morgan, 1992). The Yellowstone hot spot emerged with a burst of basaltic volcanism that may have formed the Columbia Plateau flood basalts to the northwest and the northern Nevada rift to the southeast (Zoback and Thompson, 1978). Magmatic activity on the eastern Snake River Plain was initially characterized by pulses of silicic volcanism (each of ~2- to 3-m.y. duration) progressing toward the present Yellowstone caldera. Most of the silicic volcanic rocks decrease in age to the northeast (35 to 40 mm/yr) as a result of southwestward plate migration over the plume (Christiansen and Lipman, 1972; Armstrong et al., 1975). Extensive basaltic volcanism followed, covering the plain and persisting through Holocene time (e.g., Luedke and Smith, 1983). Seismic-refraction data indicate that much of the midcrust beneath the plain was replaced by intruded basalt (e.g., Sparlin et al., 1982). The location of the Yellowstone starting plume head and the start of the plume tail is subject to discussion, because it may have encountered the subducting Juan de Fuca plate during its ascent and part of the plume-head material could have been spread westward as far as the Juan de Fuca ridge by the descending plate (R. I. Hill et al., 1992).

While the deep origin of mantle plumes is controversial, their surface expression is dramatic and widely observed. Perhaps the two most ubiquitous manifestations of mantle plumes are voluminous volcanism and a broad regional topographic swell. The typical swell associated with plumes is ~1–2 km of uplift centered across a region ~1000-2000 km in diameter (e.g., Crough, 1983; Sleep, 1990); this uplift is largely the result of isostatic compensation in the asthenosphere and lithosphere and, to a lesser extent, a dynamic pressure gradient of flow in the asthenosphere (below the limit of detection at Hawaii) (Sleep, 1990). The interpreted anatomy of a mantle plume consists of a starting plume head generated during its initial ascent through the mantle, an active plume tail that continues to flow after the starting plume head contacts the lithosphere, and ponded plume-tail material that collects at the base of the lithosphere as the active plume tail continues to flow (e.g., Griffiths and Campbell, 1991). The initial contact of the starting plume head with the lithosphere heats a broad region, because the plume head is much wider than its tail (e.g., Campbell and Griffiths, 1990; Duncan and Richards, 1991). When a hot-spot swell forms in continental lithosphere the uplifted crust may be raised far enough above the level of midplate compression to be in a state of extension, which results in rift formation centered in the swell (e.g., Crough, 1983). The magnitudes of the generated deviatoric stresses fall short of those necessary to cause complete continental breakup (Hill, 1991). The buoyancy and swell associated with the starting plume head tend to move with the lithospheric plate (Sleep, 1990; Griffiths and Campbell, 1991). The heat associated with mantle plumes tends to be contained in the mantle near the plume for a considerable time. For example, Davies (1992) concluded that the thermal anomaly associated with the Hawaiian plume has not waned during its 43-m.y. existence because the swell does not decay monotonically with distance along the volcanic chain as would be the case in a thermal decline with age. That is, the material hotter than the normal mantle adiabat remains ponded below the lithosphere and above normal asthenosphere because of the long time required for heat applied at the base of the lithosphere to conduct to the surface and because secondary convection within the thermal boundary created by plume material is inefficient.

In Section 7.5.3 it was shown that a low-density mantle underlies the northern Basin and Range province. If the Yellowstone plume behaved like the vast majority of plumes observed worldwide, then it probably had a starting head consisting of hot, lower density mantle compared to the surrounding Basin and Range asthenosphere. The emplacement of such a low-density mantle anomaly would have caused a topographic swell across the northern Basin and Range which may have in turn contributed to gravitational spreading as discussed in Section 7.6.2 and may have extended the Basin and Range province eastward along the plume track (e.g., Anders and Sleep, 1992; Pierce and Morgan, 1992). The increased thermal input from the flux of basaltic volcanism may have also stimulated extension in the vicinity of the plume by reducing the rheologic strength of the lithosphere (e.g., Anders and Sleep, 1992). The Yellowstone plume cannot be suggested as the sole cause of the latest phase of Basin and Range block faulting between 10–13 Ma, because those structures are active far south into Mexico, well out of reach of even a very large plume head. Thus if valid, the plume-head hypothesis may be more directly applicable to the relative topographic difference between the northern and southern Basin and Range.

7.7. Summary

The Basin and Range province defies generalities. There are almost always exceptions to every broad statement that can be made. For example it might be safe to suggest that the episodes of extension on low-angle normal faulting are over and that all present-day extension occurs on high-angle faults; but a seismic reflection survey in southeastern Arizona may have identified a seismogenic low-angle normal fault there (Johnson and Loy, 1992). Thus all the following broad statements and conclusions must be interpreted with the understanding that contradictions may be found at a given locality or in a specific study.

The entire western third of North America is elevated well above sea level regardless of whether it was extended or not, apparently the result of long-term subduction. Extension has widened the zone of high elevation, causing minor subsidence in the northern Basin and Range, and major subsidence, including the opening of a new ocean basin in the southern Basin and Range. Basin and Range extension is largely confined to the orogenic crust of the western Cordillera. Tectonism within the Cordillera probably weakened the lithosphere enough to enable the broad extension that has occurred there, in contrast to the more typical pattern of narrow rifting that is observed worldwide. Also contributing to the atypical breadth and elevation of the extended Basin and Range lithosphere are the time-space varying sources of extensional stresses imposed. A complicated series of events led to the extensional stresses, beginning with back-arc extension that occurred as isolated core complexes in the Pacific Northwest region during Eocene and early Oligocene time. Back-arc extension was probably also the initial style of opening in the southern Basin and Range in Mexico, beginning in middle Oligocene time. In order for a metamorphic core complex to have formed, unusual conditions such as an increased tendency for lower-crustal rocks to flow beneath them, or a change in the stress field must have existed. Both of those conditions could have been met by increased magmatism beneath the complexes, and it may be that subduction-related magmatism localized back-arc extension to those highly extended terranes in the northern Basin and Range during Eocene time.

Over-thickened lithosphere may have been weakened by the orogenic process itself, a flare-up in arc-related magmatism associated with slowing subduction, or heat transferred from the younger, hotter oceanic lithosphere that was subducted beneath it during the latest Eocene. Those factors, combined with decreased convergence at the subduction margin may have allowed the over-thickened Basin and Range lithosphere to return to a more normal thickness. However, the present-day Colorado Plateau is only collapsing very weakly along its outer margins, and is for the most part a stable piece of lithosphere. Thus it would seem that conservation of energy constraints prevent over-thickening alone from driving the Basin and Range lithosphere into the intense extension that has occurred there, unless the lithospheric mantle beneath it was substantially different from that of the rest of the western Cordillera. In other words, gravitational collapse undoubtedly contributed to some Basin and Range spreading, but outside forces had to act on the lithosphere to stretch it beyond the point of stability.

Such outside forces may have come from the divergence between the Pacific and North American plates. Up until 30 Ma the subducting Farallon plate protected the western Cordillera from the relative northwest motion of the Pacific plate. As the Farallon plate offshore of North America diminished due to continued subduction, the Pacific plate converged on North America. Asthenosphere windows have been suggested to have evolved as the remnants of the Farallon plate sank into the mantle. There is a space problem with such models because the westward drift of the Pacific plate would have quickly exposed such windows to the surface, since the North American plate could never move fast enough to cover the window if it existed at the continental margin. If instead, the Farallon plate was mechanically underplated beneath the North American continental margin as is suggested by seismic refraction studies of the margin (Howie et al., 1993; Brocher et al., 1993), then that preserved slab may have provided the link between the Pacific and North American plates that began to move together after contact (e.g., Stock and Molnar, 1988). The post-30-Ma core complex extension that began in northern Mexico and trended northward with the growing contact between the North American and Pacific plates (Fig. 7–14) then took up the divergence between the two plates. That extension took place primarily in the weak orogenic crust behind the Mesozoic batholiths clustered along the coast, but also may have occurred in the California Continental Borderlands that now lie offshore of southern California. The growing contact between the North American and Pacific plates became partially resolved along the San Andreas transform, which is not however perfectly aligned to take up all the relative motion between the two plates. Thus broad, small-magnitude extensional deformation in the Basin and Range province became necessary in order to take up the remaining relative motion, which may have prompted the latest phase of block faulting across the province. Complicating matters further, the Yellowstone plume emerged during Miocene time, and may have emplaced low-density asthenosphere beneath the northern Basin and Range province, causing its high elevation relative to the southern Basin and Range. The plume may also have made some contribution to extension through the resulting topographic swell, and may have extended the Basin and Range province to the northeast as the continent tracked over it.

Geologic and geophysical probing of the Basin and Range lithosphere indicate a province that is highly variable at the surface, while being far more uniform at depth. This uniformity may be more apparent than real, however, because of the fact that the resolution in methods applied to the deeper study of the lithosphere is quite low as compared with direct examination of the surface. The picture of the lithosphere as it emerges with the present levels of resolution (Fig. 7–15) is a brittle upper-crustal layer that extends by a variety of modes (e.g., high and low-angle faulting, magmatism) overlying a ductile lower-crustal layer that accommodates much of the surface variation in strain by flow that equalizes the thickness of the crust to an average 30 km (± 5 km). The ductile lower-crustal layer is augmented by thin sheets of magma intruded from the mantle that give it increased mobility through advected heat. Similarly, the crust-mantle boundary is interlayered mafic intrusive rocks and mantle peridotites, and has been very hot (granulite facies in the southern Basin and Range), though 300 °C of cooling may have occurred in the north-central Basin and Range lower crust. Magmatism accompanied extension, but the exact timing and relation to extensional faulting remains controversial. The uppermost mantle is a thin high-velocity zone (7.6–8.1 km/s) that is slower than the upper mantle beneath the interior continental craton. Beneath a 20–40 km-thick mantle lid lies a low-velocity (7.5–7.7 km/s) zone of mantle asthenosphere that may range from 40 to 100 km thick. The crustal and upper mantle structures are similar beneath the northern and southern Basin and Range, which implies that the 1 km difference in topographic elevation is at least partly due to a density contrast in the asthenosphere between the two sub-provinces. The chemistry of magma rising through the northern Basin and Range province is consistent with upwelling asthenosphere there, perhaps because the

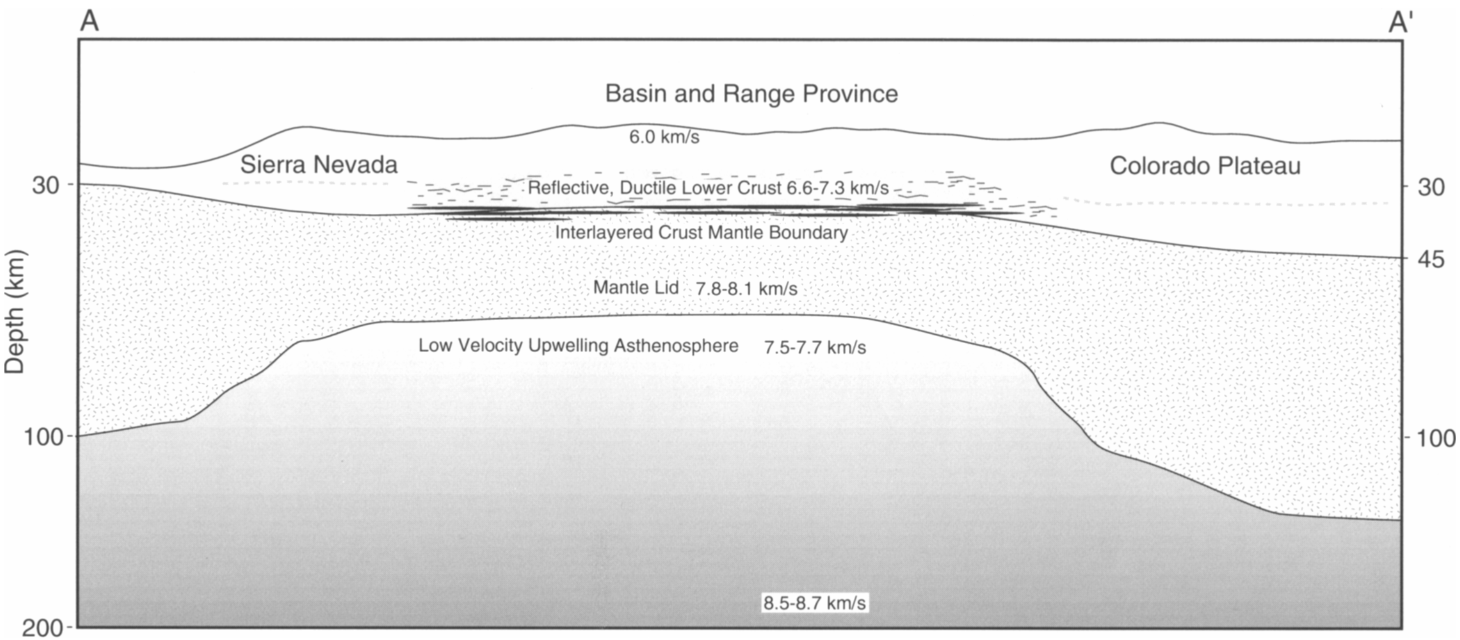

Fig. 7–15. Interpretive lithospheric cross-section (A–A' on Figure 7–1) of the northern Basin and Range province. The crust of the Basin and Range is thinner (30–35 km) than the adjacent Sierra Nevada and Colorado Plateau provinces (40–45 km), and the lower crust is highly reflective, possibly a result of ductile deformation and magmatic intrusions. The lower crustal velocities of the Basin and Range vary from about 6.6 to 7.3 km/s, perhaps because of variable magmatic underplating. Xenoliths from the crust-mantle boundary indicate that the Moho transition is interlayered magmatic intrusions and mantle peridotites. A thinned (30–50 km thick compared with 100–140 km thick beneath the Colorado Plateau and Great Valley) high velocity (7.8–8.1 km/s) mantle lid overlies a lower velocity (7.5–7.7 km/s) asthenosphere.

lithosphere has been thinned relative to the more stable regions surrounding the province, or because of active upwelling from a plume source.

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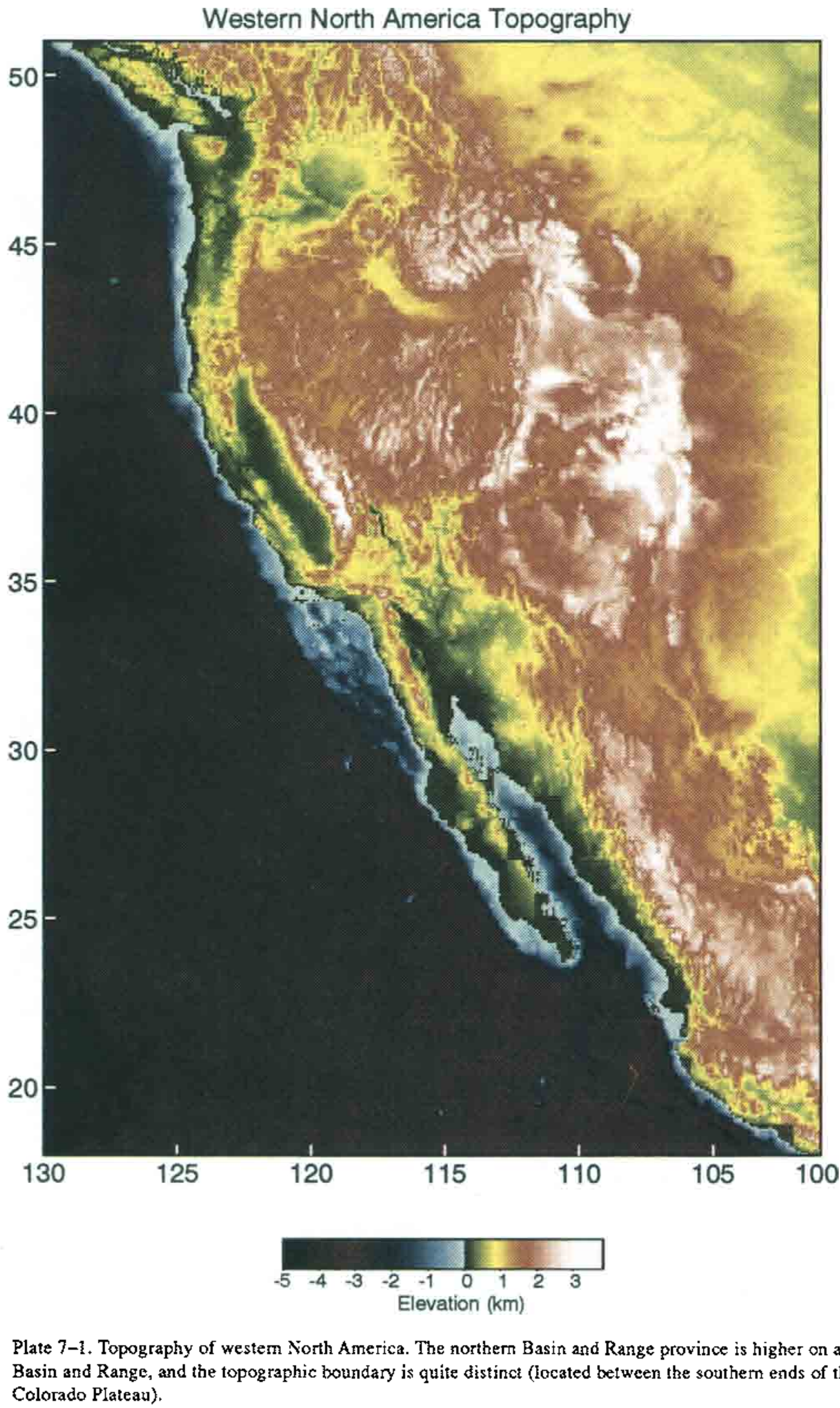

Plate 7–1. Topography of western North America. The northern Basin and Range province is higher on average than the southern Basin and Range, and the topographic boundary is quite distinct (located between the southern ends of the Sierra Nevada and Colorado Plateau).

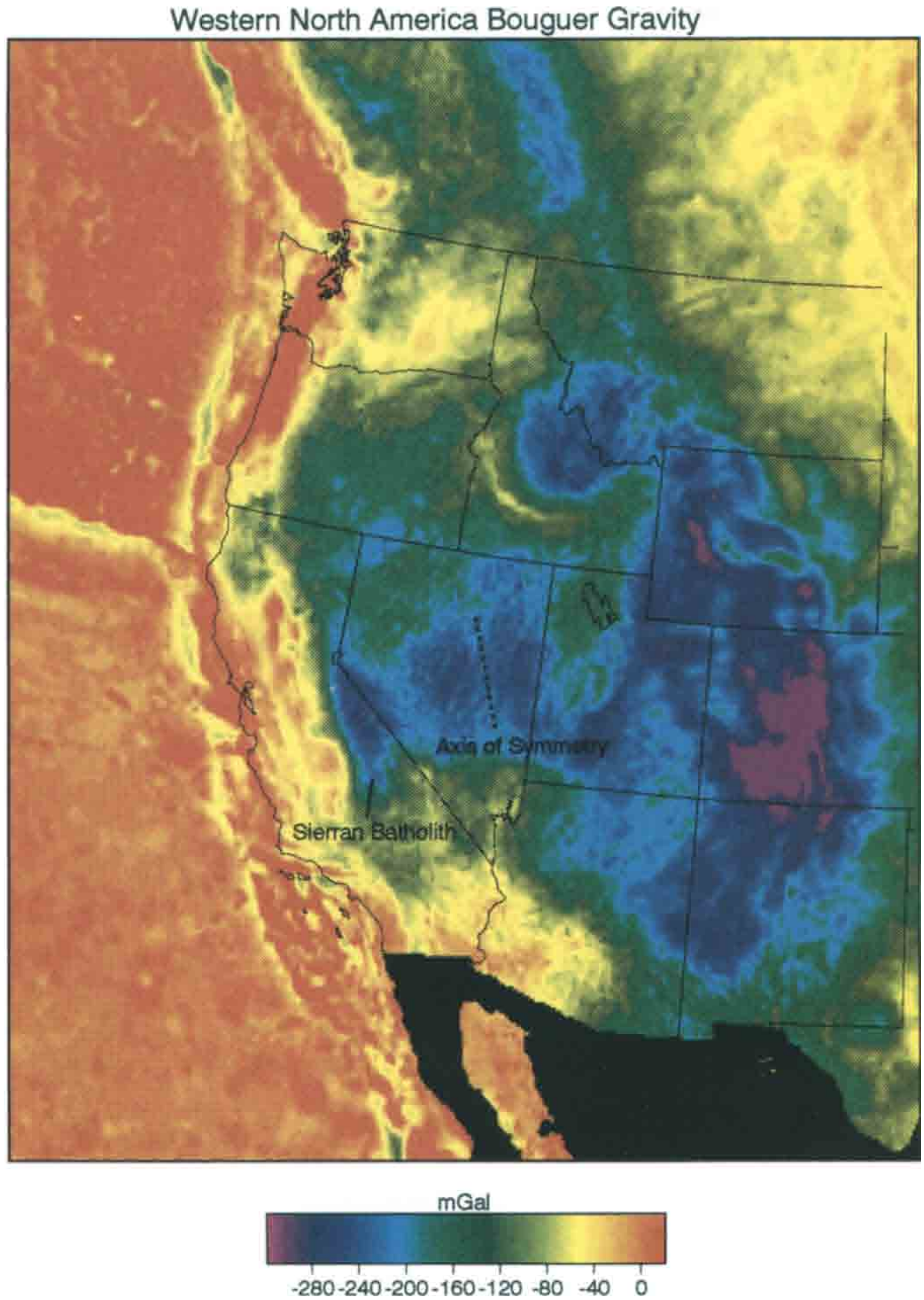

Plate 7–2. Bouguer gravity of western North America. The northern Basin and Range is a symmetric regional gravity low. See text for discussion.